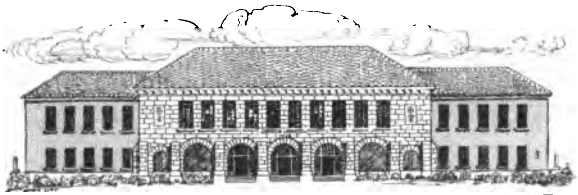

Wm Norrington

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ELEMENTARY METEOROLOGY

FOR HIGH SCHOOLS AND COLLEGES

BY

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WALDO'S METEOR.

E-P 1

PREFACE.

METEOROLOGY, which treats of our atmosphere, is a department of the physics of the globe, or "geo-physics," as it is called. It is only within recent years that meteorology has been elevated to the position of an independent science.

The more apparent atmospheric conditions have been the subject of observation and comment for many hundreds of years, but only within the past two or three centuries have accurate observations and trustworthy records been accumulated. These observations, which have greatly increased in number and accuracy during the present century, and especially during the past twenty years, are the groundwork which serves as a basis for the inductive development of the science of meteorology; and they also serve as a criterion for the testing of the truths evolved by the deductive method which has been developed by mathematical physicists almost entirely within the last quarter of a century.

It has been the custom to offer to English-reading students of meteorology the development of the subject from the inductive standpoint; and in this elementary treatise the same form has been in the main adopted.

In the deductive treatment it would be necessary to start with initial conditions, — density, temperature, moisture content, and motion of the air, — and then follow out the regular and the irregular changes of possible occurrence, keeping in mind the relation between all of these conditions and the effect of one on another. Such a treatment, while certainly more logical, and on this account more desirable, is vastly more difficult, than the method adopted, and should be offered to mature minds only, which can grasp very complex processes.

It was therefore decided to give merely the facts and their probable explanations, in treating the subject of the atmospheric conditions; furthermore, the elements have as far as possible been mentioned separately, in order to avoid the confusion of a more complex treatment, and the better to isolate the obscure and uncertain parts of the subject.

In treating of the atmospheric movements, the author has, however, combined the two methods of presentation.

The best example that we have of a non-mathematical deductive treatment of meteorology is Ferrel's "Popular Treatise on the Winds," in which that author has given the most complete account of the atmospheric circulation and allied phenomena to be found in any language, and all teachers and advanced students of meteorology are earnestly recommended to read Ferrel's book.

In order to keep this book within proper limits, the author has been obliged to omit much matter which should find place in a complete treatment of the subject. Thus but the barest mention could be made of the simpler forms of meteorological apparatus; and the application of meteorology to the sciences and arts, such as hygiene, agriculture, engineering, and to manufacturing and commerce, has been entirely excluded.

The scope of this book must not, therefore, be misinterpreted by meteorologists on the one hand, or by teachers on the other. It is intended to serve as a text-book of the elements of the science for general students, and must not be considered as a manual for practicing meteorologists.

Quantitative results have been given as far as possible in round numbers. The English systems of measurements are given, as necessary in a book intended for immediate and widespread use in our educational institutions.

In the preparation of this book the author has drawn on his experience as a former teacher of meteorology in the school of training for officers and observers of the Weather Bureau of the United States Signal Service; and he has tested much of the subject-matter herein presented (although in somewhat greater detail) in a course of lectures delivered at Evelyn College for Young Women, at Princeton.

The author is much indebted to Lieutenant Everett Hayden, U.S.N., for suggestions and assistance in reading the final proof sheets.

Many of the diagrams in the book have been redrawn from other printed sources, both foreign and American: among the latter must be mentioned in particular Ferrel's "Popular Treatise on the Winds" and Greely's "American Weather."

FRANK WALDO.

PRINCETON, N.J.

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ELEMENTARY METEOROLOGY.

CHAPTER I.

THE EARTH'S ATMOSPHERE.

Meteorology, strictly speaking, is the science which treats of the condition of the atmosphere, its changes in condition, and the causes which give rise to these conditions and changes. The science is as yet but partially developed, and much that is at present accepted as fact will be modified by future investigations.

Air. — The gaseous envelope which immediately surrounds the earth, and is called the atmosphere, is composed chiefly of air, just as the ocean is composed of water. Air is a mechanical mixture of oxygen and nitrogen gases and a small amount of carbon dioxide, together with argon and traces of various other chemical substances which have little to do with the science of meteorology as now studied. Such air is called *dry air.*

Atmospheric air always contains more or less vapor of water, which varies in amount from a very small quantity to about 5% of the amount of the dry air.

Dry Air at the earth's surface consists essentially of nearly 21% of oxygen, about 78% of nitrogen, about 1% of argon, and 0.03% of carbon dioxide, by volume.

Oxygen. — There is a slight decrease in the proportion of oxygen at distances of several miles above the sea level; and even near sea level the amount of oxygen is not absolutely invariable, but the variations are very slight. Oxygen is the most important element of the air.

Nitrogen, which forms the greater part of the air, seems to have no special function except to dilute the oxygen. It is slightly lighter than oxygen.

Argon enters into such close combination with pure nitrogen, that it has always, until its recent discovery, been included with the atmospheric nitrogen. Its peculiar function has not yet been ascertained. It is the most dense of the gaseous constituents of the air.

Carbon Dioxide, which is a compound of oxygen and carbon, varies slightly in amount, being about 3 parts in 10,000 of air on the average; but there seems to be a little more in cloudy than in clear weather, and a little more at night than by day. Carbon dioxide, though it forms such a small proportion of the air, is a very important factor in the life on the earth, since it forms the basis of food.

All animal organisms inhale oxygen, and exhale carbon dioxide. Carbon dioxide is frequently emitted as a gas from volcanoes, in such quantities as to render the air into which it passes too impure for breathing purposes, especially in low-lying adjacent places into which the gas sinks on account of its relatively great density.

Carbon dioxide is of no direct use to animals after it has been exhaled in the breathing process. It is, however, of much importance to plant life; for the carbon is the chief food of plants. The carbon dioxide of the air is taken up by certain cells of the plant, and is there decomposed by the sunlight into carbon and oxygen. The oxygen escapes

into the air, but the carbon is retained to build up the tissues of the plant.

Microscopical Impurities in the Atmospheric Air. — Minute solid particles of matter are found in the air, and these may be divided into two classes, — inorganic and organic.

The *inorganic particles* are the dust particles, which we see floating in a ray of sunlight as it crosses a darkened room; and the smoke particles, which in the aggregate are visible as smoke. The dust and smoke particles are of great importance in the changing of the water vapor in the atmosphere into drops of water.

The number of dust particles in the air varies enormously at different times and places. The number is least and most constant in clear weather, and greatest and most variable in cloudy weather; it is least over the ocean and at great distances above the surface of the earth; and the size of the particles in general decreases with the height above this surface. On the top of a high mountain, about a mile above sea level, it was found that there were 11,000 dust particles per cubic inch when the air was clear; but the number increased to 50,000 or 65,000 during the temporary fog produced by the passage of an isolated cloud over the mountain top.

The *organic particles* in the air are the minute germs called *microbes*. These are of two classes, — the *bacteria* and the *molds*. The bacteria are the minute forms of animal life, such as the disease germs; and they are most frequent in the air of hospitals, and least frequent over the oceans and on mountain tops. The molds are minute forms of vegetable life (fungi) which occasion fermentation and the decomposition of organic matter. They are most numerous in foul damp air, such as that found in sewers. Organic particles vary in number with the season of the year and the hours in the day.

Density and Volume of Gases. — Gases are elastic, and they may be compressed into a small space; but if pressure is removed, they expand, and occupy larger space: that is, the volume varies with the change of pressure. With a constant temperature, if the pressure is doubled, the volume is reduced one half. This relation is maintained for the air, within the limits of meteorological study, but for very great and very small pressures it does not hold good.

The pressure on any point in a quiet gas will be the same in all directions, — upwards, downwards, and towards either side. The air of the atmosphere is acted on by the gravity which draws it towards the center of the earth, and the amount of this downward force is the weight of the air. When we consider the outside pressures on a quantity of free air, such as may be represented by the cube shown in Fig. 1, then the pressure P_i (from without) will just equal the pressure P_r; but the upward pressure P_b will exceed the downward pressure P_a by the weight of the air, that is, by the downward pull g exerted by gravity on the cubic mass of air. Then the pressure $P_b = P_a + g$. Likewise $P_a = P_b - g$.

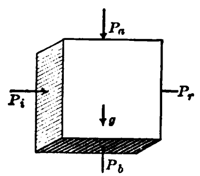

FIG. 1.

The side pressure P_i = the side pressure P_r; but these side pressures at the upper edge of the cube are each equal to P_a, and they increase, by the pressure corresponding to the weight of g, to P_b at the lower edge of the cube.

Pressure and Weight of the Air. — The weight of the atmosphere, or its downward pressure on any surface, is the result of the aggregate weight of all the air above. It is evident, then, that as we go upward from the earth's surface towards the outer limit of the atmosphere, the weight of the air, and consequently its downward pressure, will decrease; and this decrease will be in proportion to the weight of the air left below in the ascent. If the air were incompressible, then the decrease in weight and density

would be the same, for the same distance left behind in ascending, at any height in the atmosphere. The air, however, is compressible; and the greater the weight of the air above, the more the air below is compressed, and the denser it becomes. Thus, if we start from the outer limit of the atmosphere, where there is supposedly no air above, the first layer of air will be attracted by the force of gravity, but there will be no additional weight of air pressing down from above; but when we descend to the second layer of air, then gravity exerts its power on this second layer just as it did on the first, and in addition the weight of the upper layer presses down on the second layer, and compresses it, making it denser. The air of a third layer will be still more compressed, because it will have the weight of the two upper layers pressing down on it in addition to the attraction of gravity. Thus the air increases in density all the way down to the earth's surface by the compression due to the weight of the air above.

The Elastic Force of a Gas in a given quiet condition, or its expansive force, is equal to the surrounding pressure exerted on it. The different gases of the atmosphere have different elastic forces for the same quantity of matter; but for any gas the elastic force is increased if the quantity and density of the gas are increased, provided the temperature remains unchanged.

The elasticity of a gas increases as the gas grows warmer, and is measured by its pressure in the given or existing condition. For measurements of the atmospheric pressure and of the separate gases which are in the air, the pressure or weight of a column of mercury under given conditions is used; the pressure being measured by the length of the mercury column, in inches, which is necessary to counterpoise the expansive force of the gas. The instrument used

for such measurements, and called the *barometer*, is explained in the proper connection farther along.

Arrangement of the Gases forming the Air. — Each of the gases forming the air has a different degree of elasticity, and the same quantity of each occupies different volumes under similar outside pressures: so that, the greater the elastic force (pressing outward), the less will the gas be compressed for a given pressure from the outside; and consequently its density will be the less, since a smaller quantity or mass will suffice to neutralize the outside pressure.

In the case of our atmosphere, the pressure at any height above sea level is due to the weight of the mass of the gas above: consequently the elasticity or pressure, and the resulting density, decrease with the height, because, the higher up we go, the less gas there is to press down from above.

If the air were replaced by a single gas, then the greater the elasticity of this gas, the higher it would extend above the surface of the earth.

Each of the various gases forming the air has the same distribution and arrangement of its parts which it would have if it alone composed the atmosphere, so that the gases with the least density extend upward the farthest. Of the gases nitrogen, oxygen, and carbon dioxide, nitrogen is the least dense, and consequently extends up farther than oxygen, and much farther than carbon dioxide and argon, which are most dense.

It has been calculated, from our knowledge of the density of the various gases at the surface of the earth or sea level, and the rapidity with which the density of each decreases with ascent above this surface, that the carbon dioxide practically disappears from the atmosphere at a height

of about 10 miles above the sea level, the water vapor practically disappears at a height of 12 miles, the oxygen gas disappears at a height of 30 miles, and the nitrogen gas disappears at a height of 35 miles.

Observations indicate, however, that the atmosphere extends much higher than this, which shows that our knowledge of the laws of atmospheric gases is still incomplete.

Extreme Limit of the Air. — It is probable that there is no definite outer limit to the atmosphere, where we can say that the air ceases and there exists only the ether which is supposed to occupy all space. We have seen that theoretically there should be no air at an elevation of 40 miles; but meteors have been observed at a height of 100 miles or more, and these are supposed to be rendered luminous by the friction of their passage through the air.

Peculiar luminous clouds have also occasionally been observed at altitudes of upwards of 40 or 50 miles; and if, as is supposed, they are due to crystallization of water vapor, then moisture also exists at much higher elevations than the present theories will allow.

Weight of the Air. — The actual weight of a quantity of air is variable according to its density. The greater the density, that is, the more the air is compressed, the greater is its weight for any specified volume. At sea level, water weighs about 840 times as much as the same bulk of air, a cubic foot of air weighing about $1\frac{1}{4}$ ounces. With increase of height above the sea level, a cubic foot of air weighs less and less.

The Atmospheric Air Mass Relative to the Earth. — The diameter of the earth is about 7,900, and the area of its surface 197,000,000 square miles. Since the atmosphere does not, according to our most delicate tests, extend above 100 miles from the earth's surface, then its entire thick-

ness is only about $\frac{1}{40}$ of the radius of the earth; but the lower part of the atmosphere is so much denser than the upper part, that one half the total weight of the atmosphere is limited to the first three miles above the sea level, while the other half comprises the remainder of the atmosphere to its outer limit. The thickness of the lower layer, containing half the total amount of air, is, then, less than $\frac{1}{1000}$ of the radius of the earth. If the earth were represented by a globe 100 feet in diameter, this lower half of the air covering its surface would have a thickness of less than $\frac{1}{2}$ of an inch. The atmosphere, then, must be regarded as a thin gaseous layer spread out over a very large spheroid.

Distribution of the Air over the Earth. — The atmosphere has not only length and breadth, covering the whole earth's surface, but it has also thickness, extending from the earth's surface up to some unknown height. The geographical terms *latitude* and *longitude* can be used for expressing the horizontal extent of the atmosphere, or for locating a mass of air at any point on the earth's surface; but when some point above the earth's surface is to be considered, then it is necessary also to express distances in the direction of the thickness of the atmosphere, which is done by *altitudes*.

Altitude. — By the *altitude* of a place, or of a point in the atmosphere, is meant its height above the level of the ocean. Altitudes are expressed in feet, yards, or miles in the English system of measurements, and by meters or kilometers in the metric system. Altitudes are also sometimes expressed in direction merely, without reference to the absolute elevation of the point or object to be located. This is accomplished by means of angular measure, in which the altitude is the angle between a line

drawn from the observer to the object, and another line drawn from the observer to the unobstructed horizon. The angular altitude of the horizon is 0°; and of the zenith, or point directly overhead, 90°.

Importance of Altitude in Meteorology. — With increase or decrease of height above sea level, that is, with variation of altitude, the atmospheric air undergoes rapid and marked changes in its condition. At places on the ocean the altitudes at which man lives are approximately the same; but on the land this is not the case, for there we find him inhabiting regions at all altitudes from the low lands up to mountain tops reaching 15,000 feet above the sea level. Altitude thus becomes a very important matter in meteorology.

Latitude. — The heat from the sun is the chief cause of variations in atmospheric conditions: and since the average position of the sun is the plane of the equator, where the average angular altitude of the sun is greatest, and the point on the earth where the sun has the least average angular altitude is at the poles, the effect of the sun is felt most at the equator, and least at the poles; and the various gradations between these extremes are measured by any scale marking the distance between the equator and the poles. Such a measure we have in the degrees of latitude, which divide the distance into 90 equal parts. But since the degrees of latitude are counted as increasing from the equator towards the pole, — the latitude of the equator being 0°, and that of the pole being 90°, — then, as the latitude increases, the influence of the sun on the atmosphere decreases.

Longitude. — In the direction transverse to latitude, that is, in the direction in which we use longitude in geographical measurements, there is no such absolute permanent

variation in the atmospheric conditions brought about by the solar heat; and so the degrees of longitude cannot be used in expressing a law of physical change such as has just been indicated for the degrees of latitude. But still, longitude has its usual geographical significance in locating points on the earth's surface, the meteorological conditions of which it may be desired to consider.

Longitude, however, is used in another way, and that is, as representing time. We know that the rotation of the earth on its axis causes the sun to appear to travel around the earth in twenty-four hours. Now, this distance around the earth is 360° of longitude, so that the apparent motion of the sun is through 15° of longitude in one hour. The sun, then, in pursuing its apparent course in the heavens, gets back to the same position every twenty-four hours. For any one place there is thus a return to practically the same conditions, so far as the sun is concerned, at the expiration of each twenty-four hours, no matter what changes may have been experienced in the intervening hours. Such a circuit is called a *cycle.*

Meteorological Elements. — The different items by which the total meteorological condition of the atmosphere (at any place) is represented, are called the *meteorological elements.* These are, —

The *temperature* of the air, or its degree of heat.

The *pressure* of the air, or its amount or quantity, and density.

The *humidity*, or amount of water contained in the air.

The *precipitation*, or amount of water which the air loses as rain or snow.

The *evaporation*, or amount of water which the air takes up from the earth.

The *wind*, or the movement of the air.

The *clouds*, or the obscuration of the sky.

The *electrical* and *optical conditions* of the air.

Meteorological Conditions. — Observations of the meteorological elements show that they do not long remain in the same condition, but are continually undergoing changes. These changes are of two classes, — regular or periodic, and irregular or accidental.

A *periodic change* is one in which the element returns to substantially the same condition after the lapse of periods of time of approximately uniform length. The principal periods are the diurnal or daily, depending on the rotation of the earth upon its axis; and the annual or yearly, depending on the revolution of the earth about the sun.

An *accidental change* is one which occurs at irregular intervals, and whose time of occurrence cannot be foretold.

The *average condition* of any meteorological element during a given time is obtained by finding the sum of a number of single observations of its condition made during this time (usually at equal short intervals of time), and dividing this sum by the number of observations.

In addition to the average condition, it is also of interest to know the extreme fluctuations which may occur in the meteorological elements during both periodic and accidental changes. The *maximum* amount is that which the phenomenon attains when it reaches its greatest value. The *minimum* amount is that which it attains when it reaches its least value. The *amplitude* is the difference between the maximum and the minimum amounts.

Special conditions are called *phases*. The various phases of meteorological conditions are investigated as regards the time of occurrence as well as the amount or quantity.

Meteorological Instruments. — In order to observe with the required accuracy the meteorological conditions, it is

necessary to have instruments properly adapted for making measurements of the different meteorological elements. These instruments are of two classes. The first class is used by an observer for the direct observation of conditions. The second class is used for obtaining an automatic registration of conditions by mechanical or photographic means: these instruments are usually of very complicated construction.

The Distribution of Meteorological Elements is the condition of the meteorological elements at different localities during the same phases or periods of time. It is usually presented by means of geographical charts on which are entered the conditions, each at its proper locality on the map.

Statical Meteorology treats of the conditions of the meteorological elements without considering the changes in the air due to its motion of translation. It therefore embraces the whole matter of the constitution of the atmosphere, and such conditions as have the question of space, but not of time, entering into them; as, for instance, the average values of the meteorological elements for any period, or their condition at any chosen instant.

Dynamical Meteorology treats of the motions of the atmospheric air, their causes, and the conditions arising therefrom, and also of the modification which these motions cause in the statical conditions.

CHAPTER II.

TEMPERATURE.

Heat is due to very rapid motions of the minute particles, called molecules, of which bodies are composed. Heat is not a substance, but is a form of energy. It may be transferred from one body to another. Heat of various degrees or intensities may exist, and these intensities are measurable.

Temperature is a general term applied to express the intensity or degree of heat. We thus say the temperature of the body is high or low, according as the body is hot or cold. The instrument used for making accurate measurements of the temperature of bodies is called a *thermometer*, or heat measurer.

Diffusion of Heat. — The process by which heat is transferred from one body to another, or from one part to another part of the same body, is called the *diffusion of heat.* The diffusion of heat always takes place by the transference of heat from a hotter to a colder body, or from a hotter to a colder part of the same body. The diffusion of heat is accomplished by conduction, convection, radiation, or reflection, or by some combination of these.

Conduction is the flow of heat from the hotter to the colder places in an unequally heated body or in adjacent bodies. The thermal conductivity of a body is its capacity for the conduction of heat through it.

The conduction of heat does not take place with the same facility or rapidity in all bodies. The metals are good conductors, silver espe-

cially, as may be realized by putting one end of a silver spoon into hot water, and holding the other end in the hand. Air, stone, water, ice, snow, wood, and wool are poor conductors. The reason why woolen clothing is so desirable in winter is that it is a poor conductor, and so the heat of the body is retained.

Convection is the transference of the heat by the circulation of the hot body itself, by which it is brought successively in contact with colder bodies. In this process the final transference of heat from the one body to the other usually takes place by conduction. Thermal convection takes place in fluids.

Radiation is the transference of heat from a hot body through a medium which does not itself become much heated in the process, but which must be colder than the hot body. The *diathermancy* of a body is its capacity for the transmission of radiant heat without itself becoming heated.

In the passage of heat through a body, some of the heat is always retained by the body, and this retention is called *thermal absorption.*

Reflected Heat is the heat immediately thrown off from the surface of a body without entering it, when it is receiving heat by radiation from another body.

Radiant Energy. — Energy is radiated from the sun, and is transmitted to our earth by a vibratory process, through the ether which fills space. The slower vibrations are rendered sensible to us as heat, and the more rapid ones as light.

Heat of the Atmosphere. — In meteorology we have to deal principally with the temperatures at and near the earth's surface. There heat is received from two principal sources, viz., the sun, and the interior of the earth itself. The heat from the sun we call the *solar heat,* and that from the

earth *terrestrial heat.* The combined action of the solar and terrestrial heat is what we measure when we obtain the temperature of the air with which we are usually brought into actual contact. If we descend considerably below the surface of the earth, as into a mine, the influence of the heat of the earth itself is more strongly felt; while at and above the earth's surface the heat from the sun is more noticeable. The solar heat is of most importance in meteorology, and it is the solar heat which maintains the animal and vegetable life on the earth.

Solar Heat and Solar Radiation. — The sun is a very large body, about 880,000 miles in diameter, and it is intensely hot. The space around the sun is very cold, and, according to the law of the transmission or diffusion of heat, the sun must lose a portion of its heat in the endeavor to equalize its own heat and that of the surrounding space. Heat is transmitted radially (that is, straight out, like the spokes from the hub of a wheel) from the sun through this space by radiation; and, unless some matter interposes, these *solar rays* as they are called will keep on indefinitely. Our earth, in its revolution around the sun, intercepts less than one half of a millionth of the whole amount of heat given off by the sun, yet the amount received is amply sufficient for the purpose of sustaining animal and vegetable life on the earth. Although the sun is steadily parting with such vast quantities of heat, the most careful observations fail to show that it is becoming appreciably cooler.

Revolution of the Earth around the Sun. — The earth moves around the sun in one year in a slightly elliptical path or orbit, the sun being not in the center, but in one of the foci of the orbit. The earth is therefore at times slightly nearer to the sun than at other times. The

average distance of the earth from the sun is about 92,890,000 miles. On Jan. 1, when the earth is nearest the sun, it is 91,300,000 miles from it; and on July 1, when it is farthest away, it is 94,450,000 miles from it. If the earth moved around the sun in a circular path, it would receive a constant total amount of heat and light during every day of the year; but, as it is, the amount received on Jan. 1 is 7 % greater than that received on July 1.

It requires 365.2422 days, or nearly 365¼ days, for the earth to make a complete revolution around the sun. Its path lies in a plane called the *plane of the ecliptic.*

Inclination of the Earth's Axis of Rotation. — The earth has a rotation once in 24 hours around an axis passing through its center. This axis passes through the north and south poles of the earth, and is inclined at an angle of about $66\frac{1}{2}°$ to the plane of the ecliptic, or plane of revolution of the earth around the sun; so that, when the earth revolves around the sun, this axis does not stand perpendicular to this plane of revolution, but is inclined $23\frac{1}{2}°$ to this perpendicular. This is a most important matter in the formation of meteorological conditions, because it controls or marks out, to a great extent, the manner and places in which the solar rays reach the earth's surface.

Effects of the Inclination of the Earth's Axis on the Distribution of Solar Rays on the Earth's Surface. — If the earth's axis were perpendicular to the plane of the ecliptic, the length of the days and nights would be equal, and the plane of the equator would coincide with that of the ecliptic; so that during the entire year the apparent path of the sun would lie along the equator. But with the constant inclination of the axis the following conditions result during a revolution of the earth around the sun: —

The sun shines over a whole hemisphere at all times; but the part of the earth receiving the solar rays varies with the position of the earth in its orbit. The earth has the position *B* (Fig. 2) on Dec. 21, *C* on March 21, *D* on June 21, and *A* on Sept. 22. We can consider that when the sun is over the earth's equator, and consequently the earth's polar axis is turned 90° from the sun, it is in its normal or average position. On Dec. 21 the earth stands so that the north pole is turned farther away from the sun by the full inclination of the pole to the perpendicular to the ecliptic, or at an angle of $23\frac{1}{2}°$, and the sun shines on the south pole and $23\frac{1}{2}°$ beyond it (*B*). On June 21 the opposite extreme is reached; and the north pole is turned towards the sun, and the south pole away from it, by this same angle, and the sun shines on the north pole and $23\frac{1}{2}°$ beyond it (*D*). Halfway between these dates the conditions become as shown at *A* and *C*, where the region receiving the solar rays reaches from pole to pole.

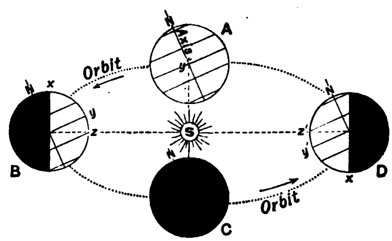

Fig. 2. — Portions of the Earth receiving the Solar Rays.

In the daily rotation of the earth when in the positions *A* and *C*, the sun appears to move through the sky in the plane of the equator. In the position *D* it appears to move in the plane of the Tropic of Cancer ($23\frac{1}{2}°$ north latitude). In the position *B* it appears to move in the plane of the Tropic of Capricorn ($23\frac{1}{2}°$ south latitude). The change from the position *A* to *B*, thence to *C*, and thence to *D*, and back again to *A*, takes place gradually, little by little each day, as the earth moves around the sun.

Periods of Presence and Absence of Solar Radiation at Places on the Earth's Surface. — Only half of the earth can receive the solar rays at any time, and, as we have just seen, the region receiving these rays varies constantly as the earth moves forward in its orbit. When the solar rays reach from pole to pole, at the points *A* and *C* (Fig. 2) the days and nights are of equal length (12 hours long) all over the world. During the time when the earth is passing from *A* to *C* through *B*, the solar rays reach beyond the south pole, but fail to reach the north pole by a like amount, and the daytime is lengthened, and the nighttime shortened, in the southern hemisphere; and for a varying region, reaching a maximum of $23\frac{1}{2}°$ at *B*, there is uninterrupted day around the south pole, and uninterrupted night around the north pole.

Similarly, when the earth is passing from *C* to *A* through *D*, the daytime is longer and the nighttime shorter in the northern hemisphere; and in a variable region, reaching a maximum at *D*, there is continual day around the north pole, and continual night around the south pole.

The days are longer than the nights in the polar hemisphere which is receiving the most solar rays; but at the equator the days and nights are always the same.

The following table shows the greatest possible length of the continued or uninterrupted visibility of the sun at various latitudes on the earth: —

Latitude	0°	17°	41°	49°	63°	66°5′	67°21′	69°51′	78°11′	90°
Hours of visibility	12^h	13^h	15^h	16^h	20^h	24^h	1 mo.	2 mo.	4 mo.	6 mo.

The twilight, which is the transition period between daylight and darkness, also increases in length with the distance from the equator. Thus, of the 8,766 hours which make up a year, there are, —

AT THE EQUATOR.	AT THE POLES.
4,407 hours day.	4,450 hours day.
864 hours twilight.	2,403 hours twilight.
3,495 hours night.	1,913 hours night.

The Apparent Motion of the Sun around the Earth. — The various phenomena just described as due to the rotation of the earth on its axis, the inclination of the axis to the ecliptic, and the revolution of the earth around the sun, are best realized by considering the matter as it appears to us.

The sun appears to revolve around the earth once in 24 hours, and on March 21 this revolution takes place in the plane of the equator; but the sun has a slow apparent movement towards the north at the same time that it moves around the earth; and the result of these two motions is a spiral movement, which carries the sun across the successive parallels, so that by June 21 it revolves around the earth on the parallel of 23½° north latitude. During this time the sun has apparently revolved around the earth 91 times. After this date there is a spiral motion southward, which gradually carries the sun back to the plane of the equator on Sept. 22. The sun then crosses the equator into the southern hemisphere, and pursues its spiral course until it reaches the plane of 23½° south latitude on Dec. 21, when it recedes again towards the equator, reaching it on March 21. As the sun thus revolves around the earth (very nearly in the planes of successive parallels), it always rises in the east, reaching the zenith at noon, and sets in the west, of places along the parallel of latitude on which it happens to be. Thus, on March 21 and Sept. 22, the sun is directly in the zenith of the equator at noon; at noon on Dec. 21 it is directly in the zenith at latitude 23½° south, and on June 21 at latitude 23½° north.

Amount of Solar Heat which would reach the Earth's Surface in the Absence of the Atmosphere. — The sun's heat which would reach the earth's surface, were there no atmosphere, would vary in intensity at different places with the angle at which the rays struck the earth's surface. When the sun's rays lay parallel with the surface, which would be the case for any special locality on the atmosphereless earth when the sun was at the horizon, the

amount of solar heat received at that locality would be zero. When the sun's rays struck the surface squarely, which would be the case when the sun reached the zenith, the intensity would reach its greatest value. In Fig. 3, the rays received between *c* and *d* are greatest, and those between *a* and *b* or *e* and *f* are least, in intensity.

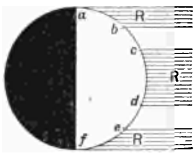

FIG. 3.—RELATIVE INTENSITY OF SOLAR RAYS.

At the time of the equinoxes, when the sun is directly over the equator, the solar rays would strike the earth's surface at the equator with their maximum strength; but at the poles they would be parallel to the surface, and therefore their effect there would be zero: and the heating power of the sun's rays would decrease, then, from the equator to the poles, or, as it is usually expressed, would decrease inversely as the latitude.

When the sun is over the equator, if the amount of solar heat at the equator were put at 1.00, then at the pole it would be 0.00. But since the sun departs from the equator towards the south 23½° at one season of the year, and a like amount towards the north at another season of the year, and since the distance of the earth from the sun does not remain a constant quantity, it becomes a very complex matter to calculate the relative amounts of heat actually received at points on the earth's surface during any 24 hours.

With the increase of latitude there is a lengthening of the number of hours a day during which the sun shines in summer, and this makes the total amount of heat received in higher latitudes during the daytime much greater than one would suppose. Thus, when the sun reaches its greatest altitude at the poles, and shines there continually, the amount of heat received there during 24 hours is over 20% more than that received during 24 hours at the equator, when the sun is above the horizon for but half the time.

If the solar radiation during the average day at the equator, which we will call the thermal day, were taken as a unit, then 365.24 would represent the annual solar radiation at the equator; and the annual solar radia-

tion at other latitudes expressed in these same thermal days would be as follows: —

Latitude	0°	10°	20°	30°	40°	50°	60°	70°	80°	90°
Thermal days	365.2	360.2	345.2	321.0	288.5	249.7	207.8	173.0	156.6	151.6

It must be distinctly remembered that these are only relative values.

The absolute quantity of solar heat which the earth would receive during the year has been calculated to be one and one half quadrillion heat units,[1] or enough to melt an ice layer over 141 feet thick, and to evaporate a layer of water nearly 18 feet deep, covering the whole earth's surface.

Solar Constant. — It has been found by experiments in recent years, that, unaffected by the atmospheric absorption and radiation, the intensity of the solar rays falling vertically is such as to heat 3 grams of water 1° C. (1.8° F.) per minute over each square centimeter of exposed surface. This number 3 is called the *solar constant.*

Solar Heat reaching the Earth's Surface in the Presence of the Atmosphere. — When the influence of the atmosphere on the solar heat is considered, the computation of the amount of heat reaching the earth's surface becomes very complicated on account of the absorption of part of the heat by the air. The amount of absorption depends principally on the length of the path of the rays through the air, the amount of moisture present, and the degree of cloudiness.

For a perfectly clear or cloudless sky when the sun is in the zenith of any point, about 75% of the solar radiation will reach the earth's surface at that point, and 25% will be absorbed by the atmosphere. But when the sun is not in the zenith, that is, when the altitude of the sun is less than 90°, then the thickness of the air layer through which the solar rays must pass will be increased, and consequently the loss of heat will be greater; while, when the sun is just

[1] The unit of heat is the amount necessary to raise the temperature of one cubic centimeter of water one degree Centigrade.

above the horizon, nearly all of the heat will be absorbed by the atmosphere, and almost none will reach the earth's surface at the point of observation.

The following table shows the proportional thickness of the air layers through which the solar rays pass at different angular altitudes of the sun, taking the vertical thickness of the atmosphere as a unit or 1.00; and it also shows in per cent the amount of the solar radiation which reaches the earth's surface, taking the amount received from the zenith on the outside of our atmosphere as 1.00.

Altitude of the sun	0°	5°	10°	20°	30°	50°	70°	90°
Thickness of the atmosphere in units	35.5	10.2	5.56	2.90	1.99	1.31	1.06	1.00
Amount of solar radiation reaching the earth	0.00	0.05	0.20	0.43	0.56	0.69	0.74	0.75

The solar rays possess different qualities, and are absorbed in different proportions in their passage through the air, according to their color, or position in the spectrum. The ultra-violet and violet rays are absorbed the most; next come the indigo, blue, green, yellow, orange, red, and ultra-red in the order named. The sun, therefore, has a more bluish appearance, the less the thickness of the atmosphere through which the solar rays pass.

The heat rays are much more readily absorbed than the light rays; so that, when the sun is near the horizon, we may receive some of the light rays, but almost none of the heat rays. The intensity of the sunlight is therefore not a direct measure of the total amount of solar radiation.

In the middle latitudes in which we live, in wholly clear weather, about half of the daily solar radiation is absorbed by the air; that is, we receive only half the amount of heat that we should receive if there were no air, or if the air had no effect on the solar heat.

The solar rays which are stopped by the air are not wholly lost to us, but we receive a portion of them by radiation from the air itself, and from the particles of matter in solid and fluid form which the air contains. But the greatest thermal use of the atmosphere is to prevent the rapid radiation of heat from the earth's surface off into space.

It has been calculated, that, if there were no atmosphere to check the radiation of heat into space, the temperature of the earth's surface would be about 325° F. below zero.

Solar Radiation on the Northern and Southern Hemispheres. — In the northern hemisphere, during the period when the sun is north of the equator, the intensity of the solar radiation is slightly less than that in the southern hemisphere when the sun is south of the equator, because in the former case the earth is at a greater distance from the sun, and the intensity of the solar rays decreases rapidly as this distance increases. The summers on the southern hemisphere are consequently slightly warmer, and the winters slightly colder, than those of the northern hemisphere, so far as the intensity of the solar heat enters into the matter. In other words, the solar climate, as the climatic influence of the sun is called, undergoes slightly more extreme changes in the southern than in the northern hemisphere. For the whole year, however, the northern and southern hemispheres receive a like amount of solar heat.

Distribution and Transference of Solar Heat on the Earth's Surface. — The solar rays which reach the surface of the earth are partly absorbed by the substance on which they fall, and partly thrown off again by reflection into the air. The effect is different for the two substances, land and water. Both of them are warmed by the addition of heat; but the rigidity of the land and the

mobility of the water, and the greater heat capacity of the latter, make the separate mention of each necessary.

The Heat received by the Land warms it very much at the surface. Part of the heat is slowly conducted by the ground to the next layers of earth below; but it does not penetrate to a depth of many feet, because the land is such a poor conductor of heat. Another part of the heat is communicated to the air layer nearest the ground, and by means of convective air currents it is transported to other localities; the change being accomplished either by the transfer of the heated air as a whole, or else by the heated air mingling with colder air and imparting heat to it. Another portion of the heat of the surface ground is lost through outward radiation into space. Another part of the heat is lost from the land by reflection. A small portion of the heat passes upward from layer to layer of air by means of conduction alone; but this is an exceedingly slow process, because air is such a poor conductor of heat.

The Heat received by the Water Surface warms it but slightly; for convection currents at once arise where the water has not everywhere the same warmth, and the heat is transferred by means of these currents to cooler portions of the water. Heat is also imparted by the water surface to the air, and is radiated outward into space as from a land surface, although more feebly, because the surface of the water is cooler; but the amount of heat returned to the air by reflection from the water surface is a very important part of the whole amount communicated to the air. Some of the heat received at the surface passes through the water for a considerable distance by continued direct radiation, and is gradually absorbed by the lower layers of the water.

A considerable quantity of heat is also used up in the process of *evaporation* from the water surface.

Principle of the Thermometer. — In general, the addition of heat to a body causes it to expand, and the loss of heat causes it to contract. Equal increments of heat cause an equal expansion, and equal losses of heat cause equal contractions, in the same body. The amount of heat added or lost can, then, be measured by the expansion or contraction which some adopted substance undergoes when subjected to these changes.

For various practical reasons, mercury is the substance chosen for the expansive or thermometric substance. It does not congeal, or become solid, in the coldest weather in most inhabited countries, and does not vaporize in the hottest weather; and, moreover, its expansion and contraction, due to small changes in the amount of heat to which it is subjected, are relatively great as compared with those of other metals, and can therefore easily be seen.

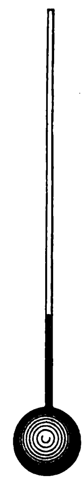

FIG. 4. — THERMOMETER BULB AND STEM.

For convenience in readily determining the amount the mercury has expanded or contracted, it is placed in a small hollow glass bulb with a connecting small bored tube as an outlet (Fig. 4). The bulb is filled with mercury at some rather low temperature, and any increase of heat will cause the mercury to flow out into the capillary tube. The larger the bulb and the smaller the bore of the tube, the greater will be the amount of rise of the mercury in the tube for a given increase in the amount of heat.

Thermometer Scales. — In order to determine just how

far the mercury moves in the thermometer tube when it is subjected to changes of temperature, it is necessary to have some kind of a scale for such measurements. If all thermometers had the same dimensions, we could use a scale of inches; but this cannot ordinarily be used, because thermometers are of various sizes.

There are several kinds of thermometer scales in use, and the divisions of the scale are called degrees, which are of the same length in each individual instrument. The scale divisions are marked either on the glass tube or *stem* of the thermometer or on a metal or glass strip placed adjacent to it.

Graduation of a Thermometer. — There are two temperatures, which, because of the comparative ease with which they may be determined, have been adopted as the basis in graduating thermometer scales. These are the temperature of freezing and that of boiling water. These temperatures are called the *fiducial points* of a thermometer. In marking the scale of a thermometer, the position of the mercury at these two points is ascertained by holding the instrument first in freezing, and then in the steam from boiling water (under normal atmospheric pressure at sea level). The graduation is then completed by marking off the distance between these points in equal subdivisions. There are three systems of thermometer graduation in use in different parts of the world, — the Fahrenheit, the Centigrade or Celsius, and the Réaumur.

In the *Fahrenheit* thermometer there are 180 divisions or degrees between the fiducial points; the freezing point is marked 32°, and the boiling point 212°.

In the *Centigrade* thermometer there are 100 divisions or degrees between the fiducial points; the freezing point is marked 0°, and the boiling point 100°.

In the *Réaumur* thermometer there are but 80 divisions or degrees between the fiducial points; the freezing point is marked 0°, and the boiling point 80°.

The graduation of standard thermometers is carried below the freezing and above the boiling points to some distance, to permit of the determination of extreme temperatures.

Temperatures below the 0° point are marked with a minus sign; those above are considered as plus, but this sign is usually omitted.

The Réaumur scale is seldom used outside of Germany at present, and is therefore not further described.

Centigrade degrees are converted into Fahrenheit degrees by the following formula: —

$$C^\circ \times 1.8 + 32^\circ = F^\circ.$$

Fahrenheit degrees are converted into Centigrade degrees as follows: —

$$\frac{F^\circ - 32^\circ}{1.8} = C^\circ.$$

Fig. 5 represents a mercurial thermometer with an attached brass scale showing degrees Fahrenheit. The temperature reading is 70°.

FIG. 5. — MERCURIAL THERMOMETER.

Effect of Change of Temperature on the Air. — Starting with the condition of the air at the temperature of freezing water (which, as we have seen, is constant, and is 32° on Fahrenheit's thermometer scale, and 0° on the Centigrade scale), the increase in volume within the limits of natural temperatures is $\frac{1}{491}$ of itself for each degree of temperature on the Fahrenheit scale, and $\frac{1}{273}$ for each degree on the Centigrade scale.

Thus, if a column of air 491 inches long at a temperature of 32° F. has its temperature raised 1° F., the air will expand to 492 inches in length; and if the temperature is lowered 1° F., it will contract to 490 inches in length. And following this law, if the temperature were lowered 491° F., then the length of the column of air would be reduced to nothing; according to which, we could not have a temperature lower than 491° F. below the freezing point of water, and this would be nearly 460° F. below zero of the Fahrenheit scale.

This assumes that the air would retain its gaseous form down to this low temperature. As a matter of fact, all of the constituents of the air would liquefy and solidify before such temperatures were reached, and so would cease to obey the law of expansion and contraction of gases, on which these results are based. Temperatures of about − 375° F. have been produced by artificial means, and air has been liquefied and solidified at a somewhat higher temperature when it has first been made very dense by compression. Of the constituents of the air, oxygen is most easily, and nitrogen least easily, liquefied and solidified.

Observations of Temperature. — Temperature observations of the air should be made in the shade, and at about six to ten feet above grass-covered earth; but in the cities the thermometers are frequently exposed from upper windows or on the roofs of buildings. Observations of temperature of the air such as these, have been made at intervals, during longer or shorter periods, at thousands of places scattered over the earth's surface. Those on the land have generally been made at fixed localities, while those on the ocean have been made mostly on moving vessels. The observed air temperatures over the globe present a wide variety of phases, which vary with the latitude, altitude, and environment of the various localities considered. Temperatures also vary with the time; that is, they do not remain constant at any one place.

Fluctuations of Temperature. — The sun in its daily course, due to the rotation of the earth, gives rise to *diurnal* fluctuations of the temperature; and in its yearly

course, due to the revolution of the earth combined with the inclination of the earth's axis, it gives rise to the gradual changes which make up the *seasonal* fluctuations of the temperature. There are also *long-period* fluctuations in the temperature, which extend over many years.

Seasonal changes in temperatures have been quite carefully studied at nearly all parts of the surface of the earth; but the changes during the day have been carefully studied at a few places only. Long-period fluctuations have been only roughly investigated.

In addition to these regular fluctuations of temperature, there are *irregular* or *accidental* fluctuations, due to causes which will be mentioned later.

The temperature at any place at the earth's surface depends principally on the length of time during which that point is exposed to direct solar radiation, and on the angle at which the sun's rays reach the place. Anything, therefore, which tends to interrupt the duration of solar radiation, or change the angle of inclination of the solar rays, must cause changes in the temperature of the point reached by those rays.

Temperature Changes during One Rotation of the Earth. — The changes of temperature during the diurnal rotation of the earth are of two classes, — the *regular* or periodic changes, due to this rotation, which vary with the different conditions under which the solar radiation is received in different parts of the earth's orbit; and the *irregular* or accidental changes, due to the oceanic or continental location, the varying degrees of cloudiness, and the movement of air masses by the winds.

These irregular changes of temperature are least in the equatorial regions, but increase towards the poles, and in middle and higher latitudes are frequently of greater

magnitude than the regular periodic changes. Their causes are mentioned later (p. 45 and elsewhere).

The Regular Diurnal Change of the Temperature is as follows: —

Beginning at sunrise, there is a gradual increase in the temperature with the increasing altitude of the sun, until the highest or maximum temperature is reached shortly after the time when the sun reaches its highest altitude in the sky; then, as the sun's altitude decreases, the temperature decreases, until (about) the time when the sun rises again to repeat the course of the previous day. There is thus a single highest and a single lowest temperature during the 24 hours.

The *maximum* or highest temperature of the 24 hours is reached at the time, after the sun has attained its greatest altitude, when the amount of heat lost from the earth by radiation just equals the amount received from the sun. This occurs at various times, ranging from 13 hours (1 P.M.) for the ocean exposures to 15 or 16 hours (3 or 4 P.M.) over the continents.

The air is radiating heat into space, and is receiving heat from the sun, during the day. When the temperature of the air is increasing, it shows that the rate at which the heat is received from the sun is greater than that at which it is lost by radiation; and it will keep on growing warmer as long as this continues, which is until a few hours after noon. But when the heat received and that given off by radiation become equal, the temperature remains stationary; and when the loss of heat is more rapid than the gain, there is a fall in the temperature.

The maximum temperature attained over the land is greater than that on the ocean; and the hour when the amount of heat received and lost is equal becomes later in proportion to this temperature, which

accounts for the earlier occurrence of the maximum over the ocean. The hour of occurrence does not vary much with the season of the year.

The degree of cloudiness — and especially during the hours about noon — greatly influences the maximum temperature. With an increase in the cloudiness, there is a decrease in the maximum temperature, because the solar rays cannot reach the earth's surface; and there is a retardation of the time of maximum, because the radiation of heat from the earth is retarded.

The *minimum* or lowest temperature of the 24 hours occurs at, or just a few minutes before, sunrise over the continents, and a little earlier than this over the ocean.

The time of occurrence is, then, at about 6 hours (6 A.M.) at the equator; and with increase of latitude, it becomes earlier in the summer half year, and later in the winter half year. In latitude 80° the minimum occurs at about 2 A.M. during the summer half year.

The minimum temperatures are lower over continents than on the oceans, because the land loses its heat more rapidly than the water. Under like conditions of surroundings, the minimum temperatures get absolutely lower with increasing distance from the equator towards the pole; but they do not descend lower below the average temperature of the place considered. This last descent is greatest at the interior of deserts in the equatorial regions.

The effect of cloud is to prevent the minimum temperatures from becoming as low as they would under a clear sky, for the radiation of heat from the earth is hindered.

The *amplitude* of regular oscillation of the diurnal temperature (or the difference between the extreme maximum and minimum during the 24 hours) is in general greatest at the equatorial regions, and decreases towards the poles, *for the same exposure;* that is, over the land or over the water. The amplitude is greatest over continents, and least over oceans, varying, in extreme cases, from about 32° F. for continental exposure in equatorial regions, to 2° or 3° F. in polar regions.

The amounts of amplitudes follow quite closely the variation in the meridian altitude of the sun and the relative length of daytime to nighttime; and consequently the amplitudes are greatest in summer, and least in winter.

The amplitudes decrease with increase of altitude above the ground, and approach in magnitude those obtaining for an oceanic exposure.

The immediate surroundings of a locality strongly affect the amount of the amplitude of daily temperature changes. If the amplitude for a plain or level surface is taken as the normal condition, then, in general, on a convex surface (a hill or mountain) the amplitude is diminished, and on a not too great concave surface (a valley) the amplitude is increased. In the case of the hill, the heat rays are dispersed, while in the valley they are collected.

The daily amplitude is greatest on clear days, and least on cloudy days, because the maximum temperature is higher, and the minimum lower, on the clear day.

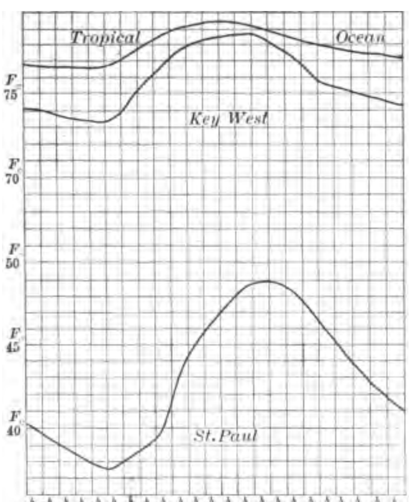

FIG. 6. — DIURNAL CHANGE OF AIR TEMPERATURE.

Representation of Diurnal Temperatures. — The daily course or march of the temperatures (and of the other elements) from hour to hour is represented by meteorologists in several ways: (1) merely by numbers, which may be only the actual temperatures, or may be the average temperature for the day, minus the individual hourly temperatures; (2) by a graphical process (Fig. 6); and (3) by means of a mathematical formula, called Bessel's Formula, which expresses this curve mathematically.

The graphical process is the simplest and best method for illustrating clearly the rise and fall of the temperature indicated by the thermometer readings. A piece of paper is divided into small, equal squares by vertical and horizontal lines. The consecutive vertical lines represent consecutive hours, commencing either at 0^h or midnight, or at 1^h; the hours are marked in a horizontal row below the diagram, at the foot of the vertical lines; and the consecutive horizontal lines represent consecutive degrees of temperature, commencing at some degree below the lowest temperature which is to be represented, and the degrees are marked in a vertical column to the left of the diagram. On the first vertical line put a dot at the proper point for showing the temperature at 0^h or 1^h, whichever is chosen to begin with, according to the scale of degrees marked on the left of the diagram; do this likewise on the next vertical line for the temperature at the next hour; and so continue for all the 24 hours. Then join these points by a succession of short lines, and the resulting curved line will show the gradual rise and fall, or increase and decrease, of the temperature for the day.

The diurnal changes in the temperature, taking the average for the year, on the island of Key West, Fla. (ocean coast), at St. Paul, Minn. (continental), at Fort Conger in the Arctic region, and also over the tropical oceans, are shown by the following hourly temperatures in degrees Fahrenheit:—

	1^h	2^h	3^h	4^h	5^h	6^h	7^h	8^h	9^h	10^h	11^h	12^h
Tropical Ocean,	77.0	76.9	76.8	*76.7*	*76.7*	76.8	77.3	77.9	78.5	79.0	79.3	**79.5**
Key West, Fla.	74.1	74.0	73.8	73.6	73.5	*73.4*	74.0	75.3	76.4	77.3	78.1	78.3
St. Paul, Minn.	40.3	39.7	39.1	38.5	37.9	*37.6*	38.1	38.9	40.6	42.5	44.5	46.2
Fort Conger,	−1.1	*−1.3*	−1.2	−1.1	−1.0	−0.7	−0.4	−0.2	0.1	0.6	1.0	1.2

	13^h	14^h	15^h	16^h	17^h	18^h	19^h	20^h	21^h	22^h	23^h	24^h
Tropical Ocean,	**79.5**	79.3	79.2	79.0	78.5	78.2	78.0	77.8	77.6	77.5	77.4	77.3
Key West, Fla.	78.5	**78.7**	**78.7**	78.2	77.6	76.7	75.7	75.3	75.0	74.8	74.6	74.4
St. Paul, Minn.	47.4	48.4	48.8	**48.9**	48.5	47.6	46.3	45.0	43.9	42.8	41.9	41.0
Fort Conger,	1.2	**1.3**	1.2	1.0	0.8	0.6	.03	0.0	−0.2	−0.5	−0.8	−0.8

The heavy-faced type shows the maximum, and the Italics the minimum, hourly temperature.

The same data, except for Fort Conger, are shown by the graphical method in Fig. 6.

The Average Daily Temperature. — This is obtained accurately by taking the sum of the temperatures observed at each of the 24 hours, and dividing it by 24. Generally, however, observations are made but two or three times a day; and it is customary to choose such hours of the day for times of observation as will best allow the average temperature to be computed from the observations made at them. Sometimes two separate hours are chosen, but more frequently three; and, in addition, sometimes the readings of maximum and minimum recording thermometers are also available.

If the sum of the temperature observations made at 7 A.M., 2 P.M., and 10 P.M., is divided by 3, very nearly the true average daily temperature will be obtained; and special combinations of other hours will give nearly as accurate results. The average of the observations at 8 A.M. and 8 P.M., as observed by the United States Weather Bureau, gives, within a fraction of a degree, the average temperature for the day.

There must be two times during the daily march of the temperature when the average temperature of the day is reached, — one during the time of increase of heat, and the other during the time of decrease, — because this average lies between the maximum and minimum temperatures. The average temperature for the day occurring during the morning increase of heat is reached at about 8^h over a water surface, and at about 9^h over a land surface; and during the evening decrease of heat, it is reached often as late as 4 hours after sunset in winter, and at about sunset in summer, over the land surface, while over the water surface it occurs at about an hour earlier than over the land.

Temperature Changes during the Revolution of the Earth. — The inclination of the earth's axis causes a difference in the length of the days and nights, and a variation in the intensity of the solar rays at points on the surface, when the earth is in different parts of its orbit.

The meteorological seasons (winter, spring, summer, and autumn), while each embraces three months in our temperate latitudes, are quite arbitrarily fixed when we consider the earth's surface as a whole; for they neither occur during the same months in the two hemispheres, nor do they have the same duration in different latitudes. With us the coldest month (January) is taken as the mid-winter month, and the warmest month (July) as the mid-summer month; and the mid-spring month (April) and the mid-autumn month (October) come midway between these. It is only in middle latitudes that the four seasons are all well marked and of approximately equal length. In lower latitudes the change from winter to summer is but slight, and the transition occurs so gradually as to be almost imperceptible so far as the temperature is concerned. In the very high latitudes, on the contrary, the change from summer to winter and from winter to summer is very abrupt, and there is really only a winter and a summer season.

The seasons, except for local purposes, are to a great extent losing their former importance in meteorological reports, because of late years nearly all data are given for the separate months and for the calendar year, in order that the observations made all over the globe may be readily compared. For agricultural purposes the average meteorological conditions during the seasons are important.

The times of the seasons are reversed in the northern and southern hemispheres. When it is winter in the northern hemisphere, it is summer in the southern hemisphere; and when it is spring in the northern hemisphere, it is autumn in the southern hemisphere. In middle latitudes, where three months are given to each season, they are as follows: —

Months.	Dec. Jan. Feb.	Mar. Apr. May.	June. July. Aug.	Sept. Oct. Nov.
Seasons in the northern hemisphere.	Winter.	Spring.	Summer.	Fall.
Seasons in the southern hemisphere.	Summer.	Fall.	Winter.	Spring.
In Fig. 2 these periods are shown at	B.	C.	D.	A.

Gradual Change of Seasons. — The change in the angle of inclination of the solar rays, and in the length of days, from day to day during the progress of seasons, is very slight; and except near the poles the resulting change in the temperatures of places on the earth's surface is correspondingly small. We have, then, slow temperature changes during each season, and but a gradual passing from one season to the next.

In following out the changes in temperature due to the revolution of the earth around the sun, we have, then, naturally to turn to the series of average temperatures for each day for successive days during the year. Accidental causes which give rise to variable temperature conditions from day to day, so disguise this slow gradual change, that we can realize its occurrence only after the lapse of a number of days.

The Yearly Course or March of the Temperature. — This is represented by the series of average temperatures for equal fractions of a year, usually for the months; but in some cases five-day averages are used. Single daily averages are not used, because even a century of observations would not give the average daily temperature with sufficient accuracy to remove all accidental irregularities which occur in them.

There exists during the year (away from the equatorial

region) but a single maximum and a single minimum, as was seen to be the case for the daily temperature. In the northern hemisphere the warmest month is July, and the coldest is January, which is about a month later than the times when the sun reaches the highest and lowest altitudes respectively. On and near the oceans the greatest retardation takes place. In the southern hemisphere the months of maximum and minimum temperatures just stated are reversed. It thus occurs that in the region about the equator, the maximum temperatures occur at the time when higher latitudes are enjoying medium temperatures (spring and fall). Two maxima and two minima thus exist for the equatorial region, but the amplitudes are comparatively slight.

The reason that the highest and lowest monthly temperatures occur later than the times of highest and lowest meridian altitude of the sun, is similar to that for the retardation in these phases in the diurnal changes of temperature. When the whole amount of heat received from the sun during the 24 hours of the day just equals the amount lost from the earth by radiation, then the average daily temperature remains the same. But as the sun gets higher in the heavens in the spring, the amount of heat received from it exceeds more and more the amount lost by radiation, and the days grow warmer. When the sun recedes from the tropics in summer, the amount of heat received still exceeds for another month or so the amount lost: consequently the average temperature still increases during this time, and continues to until the sun has moved so far equatorward that the amounts of heat received and lost are equal. The average temperature also continues to decrease during some weeks after the sun has reached its lowest meridian altitude, for the amount of heat lost by radiation still exceeds the amount received from the sun until the sun's altitude has considerably increased. In our latitude the two become equal in January.

The *amplitude of the yearly temperature period* is usually understood to be the difference between the

highest (maximum) and the lowest (minimum) *monthly* averages. This amplitude, for the same surroundings, increases from the equator poleward. The land and water distribution very powerfully influences the yearly amplitudes, which are least over the oceans; so that a very dry continental location near the equator may have as great an amplitude as a moist oceanic location far to the poleward.

For instance, the yearly amplitude at Biskra in the Sahara (continental) is 39.8° F., and at Madeira (oceanic) it is but 10.8° F. At Vardo (oceanic), northwestern Europe, the amplitude is 25.9° F., and at Werchojansk (continental), Siberia, it is 117.9° F. At this last place there has been observed a temperature of −90° F. in winter, and of +86° F. in summer, thus making an absolute extreme amplitude of about 180° F. in the year.

The temperature amplitudes for the year in our middle latitudes increase along the same parallel from the western coast towards the middle and eastern parts of the continents, and then decrease again towards the eastern coasts. On the eastern coast the amplitudes are greater than on the western coast, on account of the great influences of the prevailing winds.

With increase of altitude, the yearly amplitude decreases, and more nearly approaches that for a marine exposure at the same latitude.

Below are given, in degrees Fahrenheit, average monthly temperatures for Key West, Fla., St. Paul, Minn., and Fort Conger in the Arctic region.

	Jan.	Feb.	Mar.	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Key West (Seashore) . . .	69.7	71.4	73.2	76.6	79.8	83.0	84.1	84.3	82.2	78.8	74.3	70.0
St. Paul (Continental) . .	11.7	17.5	28.2	44.9	58.8	67.2	71.9	69.2	58.7	47.0	31.0	18.1
Fort Conger . . .	−38.2	−40.1	−28.1	−13.6	14.1	32.6	37.1	33.8	15.8	−8.9	−23.6	−28.1

The same, except for Fort Conger, are shown graphically in Fig. 7.

Excessive Range of Monthly Temperatures. — Excessive amplitudes of temperature are due to the extremely low minimum temperatures, and not to unusually high maximum temperatures; the minima going lower than usual, and not the maxima going higher.

This is realized by considering the fact that we should call a day warm when the thermometer reaches 85° F., but very hot if it reaches 100° F., the difference being but 15° F. While 30° F. would be considered quite cold, yet a fall of the temperature to 20° F. below zero is not an unusual experience in the northern part of the United States; and this would mean a lowering of the temperature by 50°, which is over three times as great as the increase of 15° F. just mentioned.

The highest average monthly temperature ever observed is that of 102° F. for July, at Death Valley, Cal.; and the lowest is −60° F. for January, at Werchojansk, Siberia. Probably the temperatures in some parts of the Desert of Sahara are greater than those for California, but we have little accurate knowledge of the temperatures of that African region.

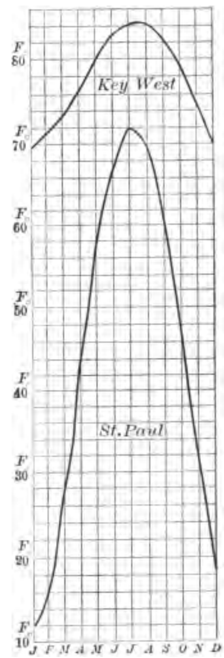

Fig. 7. — Annual Change of Air Temperatures.

The *average temperature for the year* is obtained by taking the sum of the average temperatures for all the months, and dividing it by the number of months. Since monthly temperatures vary, so the yearly average temperature will vary, but not so much as the monthly temperatures; and these last, in turn, are less variable than the daily temperatures.

Irregular Changes of Temperature occur, which depend on the degree of cloudiness, effects of radiation, and on the

movement of air masses. These temperature changes are local, but may be of wide extent. When irregular rapid fluctuations of the temperature take place, the distribution of the air temperature near the earth's surface is usually as follows: —

There is a central, more or less limited, area of high or low temperature, from which the temperature irregularly but gradually decreases or increases, as the case may be, in all directions.

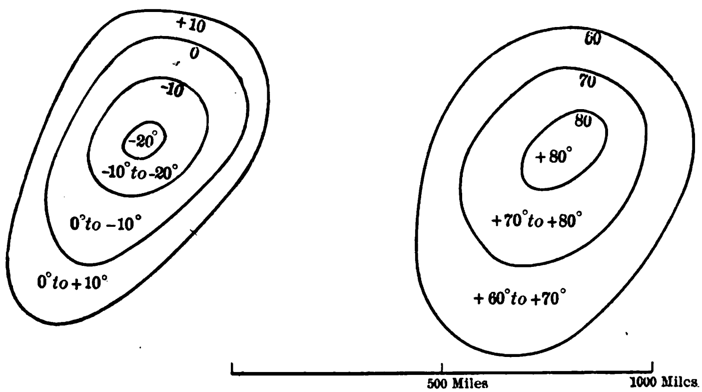

FIG. 8. — AREAS OF LOW AND HIGH TEMPERATURE.

These areas are shown in a general way in the accompanying diagrams (Fig. 8), in which the curved lines inclose limited areas on the earth's surface. In the figure on the left, the air in the central area has a temperature of $-20°$ F. or less; around this is a zone in which the temperature is from $-20°$ to $-10°$ F., and around this are others in which the temperatures are from $-10°$ to $0°$ and from $0°$ to $+10°$ F. respectively. In the figure on the right, the central area has a temperature of 80° F. or over, and within the successive surrounding rings the temperature is from $+80°$ to $+70°$ F., from $+70°$ to $+60°$ F., etc. Many times these areas do not appear closed on all sides, and this regularity of increase or decrease of temperature does not take place on the open side within the limits of the region considered. These

areas of abnormally cold or warm air move over the earth's surface in regular tracks, and this causes the temperature irregularities for particular places to have quite constant average values, depending on the location of these places with reference to the tracks.

Sometimes these areas extend over many thousands of square miles, the inner region of greatest or least temperatures sometimes covering more than one of our States.

The Temperature Anomaly. — If the average temperature for the same month during many successive years is obtained, and the difference between this average and the monthly temperature for each year is taken, the mean of these differences will be the *mean temperature anomaly* for that month. Temperature anomalies are the least in lower latitudes, and are greater at the interior of continents than on the oceans. They are the greatest in winter, and least in midsummer. The mean anomalies are relatively quite constant for adjacent places.

Variability of Temperature. — The variability of temperature from day to day is the difference between the average temperatures for successive days. The average of such temperature differences is usually computed for each month. Temperature variability is least in the tropics, and increases with the latitude up to about latitude 50°, whence it decreases again toward the poles. It increases from the sea towards the interior of continents, and also with the altitude on the average for the year; and up to a certain limit it decreases in summer and increases in winter with the altitude. It is greater on the east than on the west coast in our latitudes, and is greatest in winter and least in summer.

Concerning the extreme variability of mean daily temperature from day to day, it may be said that changes of 4° F. occur in all parts of the world, of 7° F. in most parts, and of 10° F. in but few parts. In the interior of

continents, as, for instance, North America and Asia, individual changes of even 45° F. occur, but at rare intervals. The amount and frequency of the individual temperature changes increase towards the interior of a continent, but in our middle latitudes are greater on the eastern coast of the continent than on the western.

Average Temperatures for Adjacent Regions. — The difference in the temperature for adjacent places is a much more constant quantity than are the absolute temperatures themselves. For instance, suppose that the temperature has been observed at two neighboring places, — at the one for the last 100 years, and at the other for only the last 10 years, — then the best way of obtaining the true average temperature of the second place is to find the average temperature at both places during the same 10 years, and take the difference between the two; and this difference applied to the average for the one place for 100 years will give the average temperature for the second place much more accurately than it has been determined from the 10 years' observations alone.

Long-period Temperature Oscillations are shown by direct observation of thermometers, and by several indirect methods, such as the dates of the harvests, the opening and closing of navigation by ice, the relative severity of winters, and other methods. Observations (mostly European) show a periodic fluctuation of temperature during a period of about 35 years. During the past century the years 1791 to 1805 were relatively warm; 1806 to 1820, cold; 1821 to 1835, warm; 1836 to 1850, cold; 1851 to 1870, warm; and 1871 to 1885, cold. The average fluctuation amounts to about 2° F. during these periods.

Change of Temperature with the Altitude above the Surface of the Earth. — The normal condition of the temperature of the air is a decrease with the altitude above the earth's surface. This decrease is not the same in amount

at all places on the earth, nor is it constant at any one place. In fact, it frequently occurs that for a short time, and for a short distance upward, there is an increase of temperature with the altitude. The observed decrease of temperature with the altitude is, on the average, about 1° F. for 330 feet; but it is more rapid in summer than in winter, and it is more rapid in the lower air layers than in the higher. In the winter and in the nighttime, in clear weather, the air in the valleys is cooler than that on the low hilltops; and in summer and in the daytime, the air in the low valleys is warmer.

Observations have been made on mountains and in balloons with the view of determining accurately the law of the decrease in temperature with the altitude; but such a great variety of results have been obtained, that the matter is at present by no means definitely settled.

The laws of thermodynamics and adiabatic[1] cooling show that for dry air there will be a decrease of temperature of about 1° F. for 183 feet of increase in altitude, and a corresponding increase of temperature for a like decrease in altitude. The usually observed decrease of temperature with the altitude, as has been stated, falls short of this amount. The outer limit of the decrease of temperature in the atmosphere, or the

[1] By adiabatic heating or cooling of a mass of air is meant a change in its temperature brought about by a change in its density, and without the addition of the heat from or the loss to outside sources. Dry air will cool adiabatically about 1° F. if its density is decreased by a change in pressure equal to that involved in a change in altitude of 183 feet. Hence, as a result of the decrease in the pressure and density of dry air with altitude, there will be a corresponding decrease in the temperature of 1° F. for each 183 feet of altitude. When the pressure upon a body of air is decreased, the air expands. To effect this expansion, part of the heat energy of the air is utilized, and becomes insensible as heat; hence the air becomes cooler, or its temperature falls, as a result of its expansion. On the other hand, when the pressure on a body of air increases, the air is compressed into less space; and part of the energy which was employed in maintaining its former bulk, being relieved of this duty, appears as sensible heat; wherefore the air becomes warmer, or its temperature rises, as it grows denser under compression.

temperature at points above the earth's surface where the air is very rare, is not known with any degree of certainty; but it has been thought to be about $-45.5°$ F., judging from the observed decrease of temperature up as high as ten or twelve thousand feet. A temperature of $-54°$ F. has, however, been observed in a balloon at an altitude of 31,500 feet.

This temperature would be the same above the pole and above the equator. It is seen, then, that in Siberia, with an average temperature for January of $-58°$ F., there must be an increase in the temperature with the altitude for a relatively long time; but it is probable that this increase in such cases extends upwards for a short distance, and that it reaches a limit at a not great altitude, from which point there is a decrease farther upwards; that is, there is a relatively warmer mass of air somewhere between the very cold air layer at the earth's surface and the very cold outer limit of the atmosphere.

The rate of decrease varies for different months of the year, and the amount of this variation is much greater for inland mountain regions than for those near the ocean. The maximum rate of decrease is about 1° F. for 200 feet in most cases; and the great variation which occurs for individual cases is due to inequalities in the minimum rates, which appear to increase with the latitude. The average rate of decrease for the year, while varying from 1° F. per 370 feet to 1° F. per 220 feet, is probably about 1° F. in 330 feet altitude.

The rate of temperature decrease with the altitude also varies from hour to hour during the day, and moreover is greatly influenced by the degree of cloudiness.

Near the ground at midday in clear weather the decrease of temperature with altitude is much more rapid than in cloudy weather; but at evening the decrease in cloudy weather is greatest.

At an altitude of several hundred feet above the ground the midday decrease in clear weather is not much greater than for cloudy weather, and may even become equal to it; but in the evening the decrease is less during clear weather than for cloudy weather. The decrease of temperature in cloudy weather is, then, quite constant for both day and night above the very lowest air layers; but in clear weather it varies greatly. The cases of inversions of temperatures, or increase of temperature with altitude, occur in clear weather.

Geographical Distribution of Air Temperatures over the Earth's Surface. — If the earth had a homogeneous surface

of either land or water, the average temperature for the year would decrease from a maximum at the equator to a minimum at the poles, and the average temperature along any parallel could be considered as dependent on the latitude alone.

If the surface were all land, it would have a higher temperature than if it were all water, in the equatorial regions, because land is more easily warmed than water; but in the polar regions the water surface would have the higher temperature, because the land loses its heat by radiation faster than the water. At some middle latitude the temperatures for the two surfaces would be equal.

If we compare the temperatures for an ideal land-covered with that of a water-covered earth, then at the equator the temperature of the land would be 115° F., and of the water nearly 72° F.; and the difference between the equatorial and polar temperatures would be, for the land about 135° F., and for the water about 54° F.

In the actual case of the unequal and irregular distribution of land and water on the surface of the earth, however, the average temperature along any parallel is dependent on this distribution as well as on the latitude. There are two other causes which are also effective, but to a much less degree. These are the transference of heat toward the poles, and cold toward the equator, by means of ocean currents; and the transference of heated or cooled air by the prevailing winds. The effects of these, however, are practically neutralized over the entire circumference of the earth, because where one part is rendered cooler by a cold current, another is made warmer by a warm current.

Isothermal Charts. — An *isothermal line*, or *isotherm*, is a line every point of which has the same temperature. An *isothermal surface* is a surface every point of which

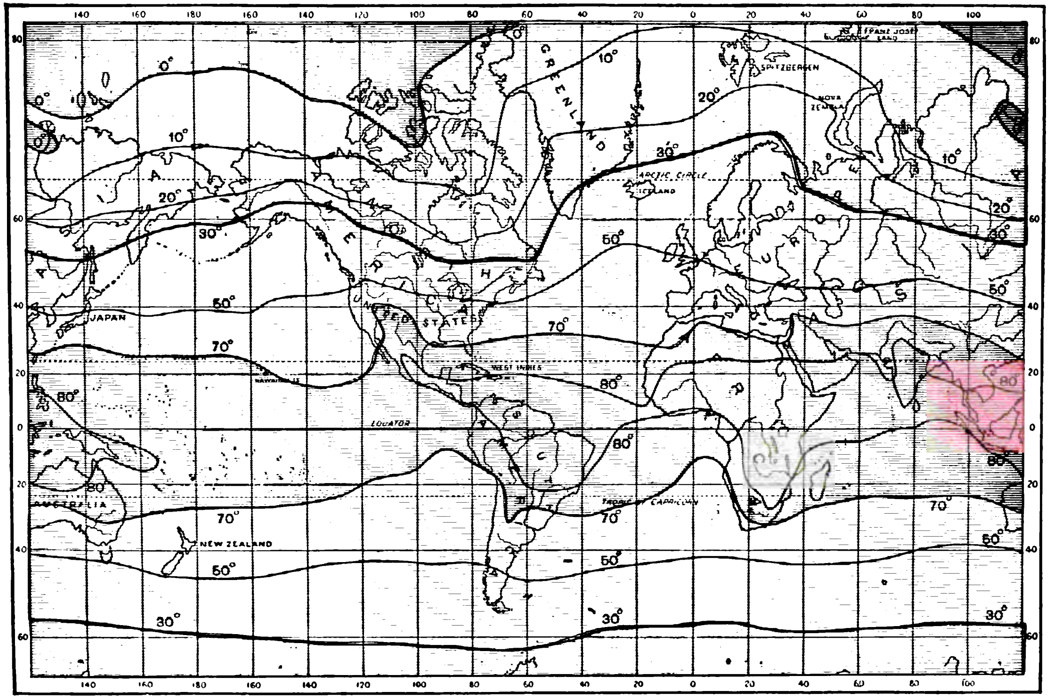

FIG. 9.—AVERAGE TEMPERATURE FOR THE YEAR (AFTER BUCHAN).

has the same temperature. When the average temperatures for the year at different places are written down on a map at those places, and the isothermal lines are drawn on the map, then we have a chart of the isothermal lines for the year for the region covered. These isothermal lines show where the isothermal surfaces of the air, which are inclined downwards from the equatorial towards the polar regions, intersect the level of the surface of the earth. Charts of this kind have been made for the whole earth, except for the polar regions, for which we have few observations. Similar charts have been constructed, showing the average temperatures for each month of the year. Such charts for the year (Fig. 9), for the midwinter month, January (Fig. 10), and for the midsummer month, July (Fig. 11), are given here.

Courses of the Isotherms. — As was to have been expected from what has just been said, the isothermal lines for the year (Fig. 9) do not follow the parallels of latitude around the globe. These irregularities are due to three main causes: —

1. The unequal distribution of the land and water surfaces, and the excessive heating of the land in summer and cooling in winter, as compared with a water surface.

The irregularities are much more marked in the northern than in the southern hemisphere, because in the former the ratio of land to water is the greater. The excessive cooling of the interior of the continents in winter causes the isothermal lines to make there a bend equatorward at that season, while in summer the excessive heating of the land causes the isotherms to make a poleward bend.

2. The transference of the equatorial heat poleward, and the polar cold equatorward, by means of the oceanic currents; and this heat and cold are communicated to the air.

The great oceanic currents give a greater warmth to the waters in the western parts than to those in the eastern parts of the oceans up to about latitude 40°, where the warm current coming from the equator, along the eastern coast of the continent, crosses the ocean; and, dividing as it approaches the western continental coast, part of it flows poleward, taking abnormally warm water with it, while part of it flows equatorward, taking with it colder water than was to be found at the same latitude on the western side of the ocean. On the eastern continental coast to the poleward of latitude 40°, where the warm current from the equator crosses the ocean, there is a cold current coming from the polar regions, which makes the coast waters abnormally cold.

3. The transference of heated and cooled air by the prevailing winds.

In the middle latitudes of the two hemispheres the winds from the west blow in winter the colder air from the continents on to the western parts of the ocean; and the isothermal lines crossing the oceans from west to east are inclined poleward (that is, in a northeasterly direction in the northern hemisphere), and separate farther and farther apart in their progress. The isothermal lines crossing the continents are inclined toward the equator (that is, in a southeasterly direction in the northern hemisphere), and converge more and more in their progress, because the prevailing winds are from the west, and in winter blow the warmer oceanic air on to the western coasts of the continents, and the cooler continental air on to the eastern coast regions. In the summer the winds from the west blow the cooler oceanic air on to the western coasts of the continents, while the same winds blow the heated continental air on to the eastern coasts. Thus the western coasts are kept cooler, and the eastern become warmer.

In the equatorial zone the winds blow mostly from the east, and the conditions depending on the general direction of the wind are in general the reverse of those in middle latitudes.

A more detailed statement of these reasons is deferred to the chapter on climates.

Normal Average Temperatures. — The average temperatures for the different degrees of latitude have been derived

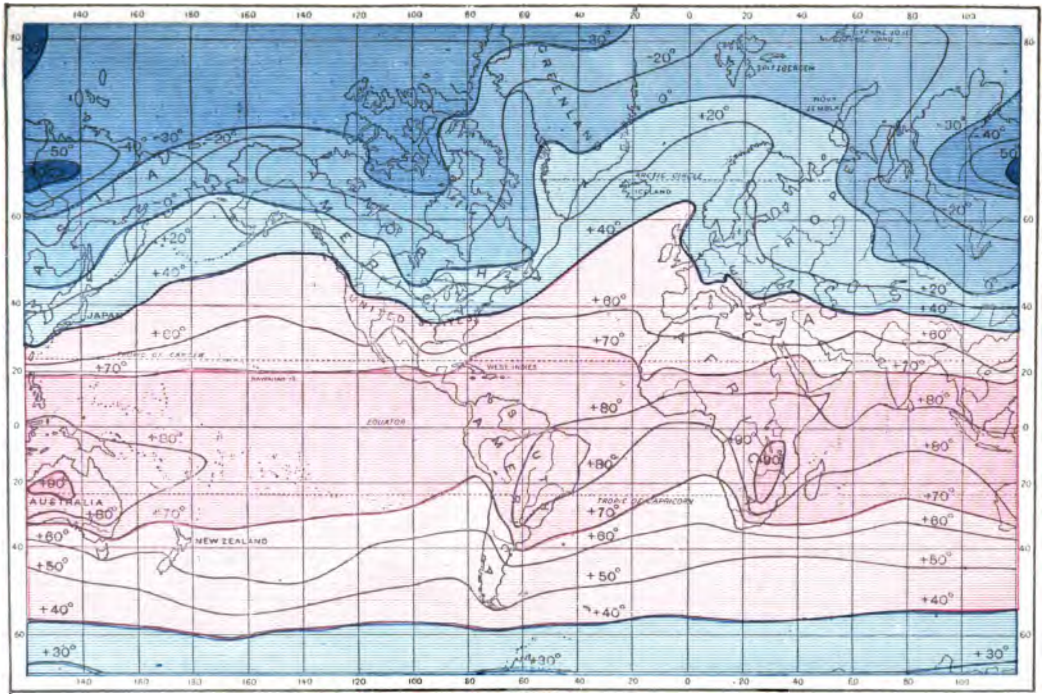

FIG. 10. — AVERAGE TEMPERATURE FOR JANUARY (AFTER BUCHAN).

by taking the average of the temperatures at various points along individual parallels. The following table shows these average temperatures for the year, for January, and for July, along each 10° of latitude, both north and south of the equator: —

LATITUDE.	JANUARY.	JULY.	YEAR.
	F.	F.	F.
North 80°	−30.9°	+32.3°	+ 1.6°
North 70°	−15.7°	+44.1°	13.7°
North 60°	+ 3.9°	56.9°	29.8°
North 50°	19.9°	64.6°	42.5°
North 40°	42.9°	75.4°	57.1°
North 30°	59.5°	81.0°	68.4°
North 20°	71.7°	82.4°	76.9°
North 10°	78.5°	80.7°	80.8°
Equator 0°	80.0°	78.2°	79.9°
South 10°	80.0°	74.9°	78.2°
South 20°	77.6°	66.9°	73.9°
South 30°	70.0°	57.1°	64.0°
South 40°	59.1°	46.9°	54.0°
South 50°	47.4°	+36.8°	+41.5°
South 60°	+34.9°	——	——

Up to about 45° latitude the northern hemisphere is warmer than the southern, but beyond that latitude the southern hemisphere is the warmer. The warmest parallel for the year, or the thermal equator, is found to be about 10° north of the equator.

The rate of poleward decrease in the temperature is irregular, since the relative amounts of land and water vary. The differences between the midwinter and midsummer temperatures for the two hemispheres are found to be very unequal; but the average for the January and July temperatures gives nearly the annual temperature in the respective hemispheres.

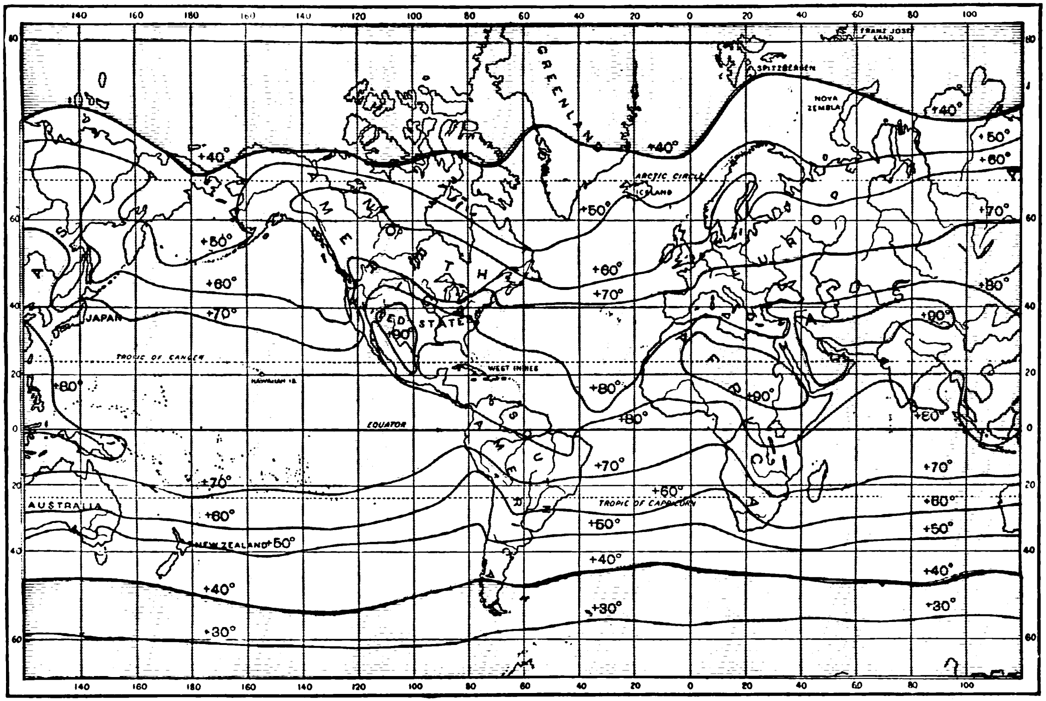

FIG. 11.—AVERAGE TEMPERATURE FOR JULY (AFTER BUCHAN).

The average surface air temperature of both the northern and southern hemispheres is about 59° F. For the extreme months January and July the average temperatures have been computed as follows: —

	January.	July.
	F.	F.
Northern hemisphere	46.4°	72.5°
Southern hemisphere	63.5°	54.3°
Whole earth	55.0°	63.3°

It has also been found, that, on account of the greater amount of land in the eastern hemisphere (counting this from 80° west longitude to 100° east longitude from Greenwich), the eastern hemisphere is about 2° F. warmer than the western.

Abnormal Average Temperatures. — The irregular distribution of temperatures over the earth is best shown by considering the difference between the average temperature of each parallel and the actual temperature at places along the same parallel. These differences are called the *abnormal temperatures* of those places. The abnormal temperatures are charted, and the lines of equal magnitudes drawn on the charts are called *is-abnormals*.

For the whole year and for the winter the temperatures are abnormally highest over the eastern parts of the oceans, and abnormally lowest over the centers of the continents in northern latitudes.

For the summer the temperatures are abnormally highest over the centers of the continents, and abnormally lowest over the eastern parts of the oceans in northern latitudes. Such charts are shown, for January in Fig. 12, and for July in Fig. 13.

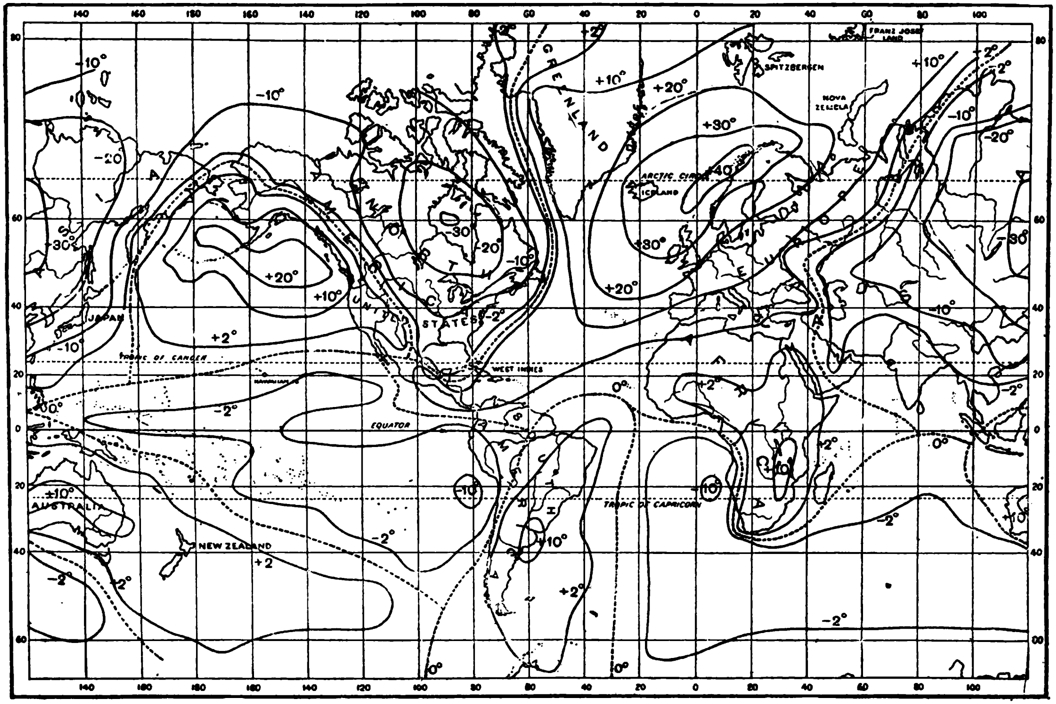

FIG. 12.—IS-ABNORMAL TEMPERATURES FOR JANUARY (AFTER BATCHELDER).

Abnormal Temperatures for the Year. — The regions where the local temperatures depart most below the normal are in north-central North America (−12° F.) and northeastern Asia (−18° F.). The regions where the local temperatures depart most above the normal are off the coast of Norway (+20° F.), around the eastern end of the Mediterranean Sea (+10° F.), and on the southern Alaskan coast (+8° F.).

Abnormal Temperatures in January. — In the chart (Fig. 12), the minimum negative abnormal temperatures are −30° F. in northeastern Asia, between −20° and −30° F. in north-central North America, and −10° F. to the west of the middle latitudes of South America and the southern latitudes of Africa. The maximum positive abnormal temperatures are +40° F. off the western coast of Norway, +20° F. off the southern coast of Alaska, and +10° F. in central Australia, southern South America, and southern Africa.

Abnormal Temperatures in July. — In the chart (Fig. 13), the minimum negative abnormal temperatures are −10° F. on the northeastern part of the Pacific Ocean, on the North Atlantic between southern Greenland and the North American Continent, and to the west of South America and Africa a few degrees south of the equator. The maximum positive abnormal temperatures are +20° F. in the northwestern United States (inland), +10° F. in central Asia, and +10° F. over the desert of northern Africa. In the southern hemisphere there is a slight negative abnormal temperature in central Australia. Over the greater parts of South America and Africa there is a slight positive abnormal temperature both in winter and summer, while over the ocean to the west of these the air is abnormally cool.

Annual Range of Average Monthly Temperatures. — The difference in the average temperatures for the coldest and warmest month is called the *range* in the monthly temperatures for the year. It increases with the latitude and towards the interior of continents, but decreases with the altitude.

These temperature differences for many localities have been deduced and entered on a chart, and the lines of equal range of temperature have been drawn by connecting adjacent places having the equal differences of tem-

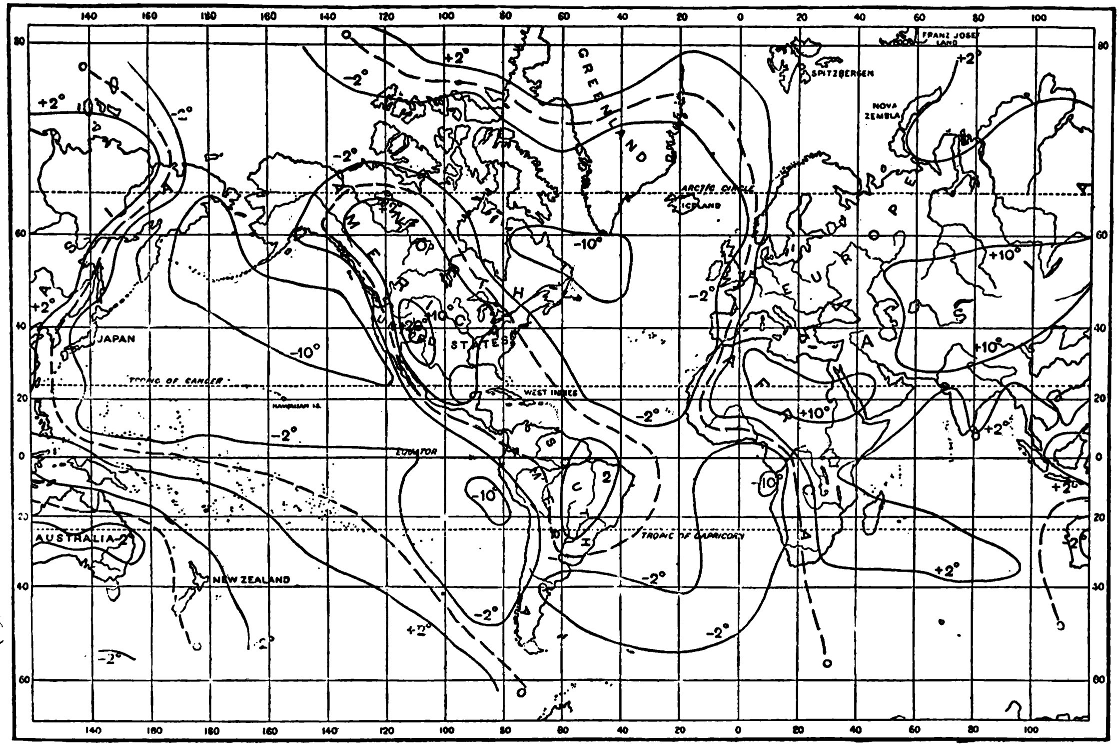

Fig. 13.—Is-abnormal Temperatures for July (after Batchelder).

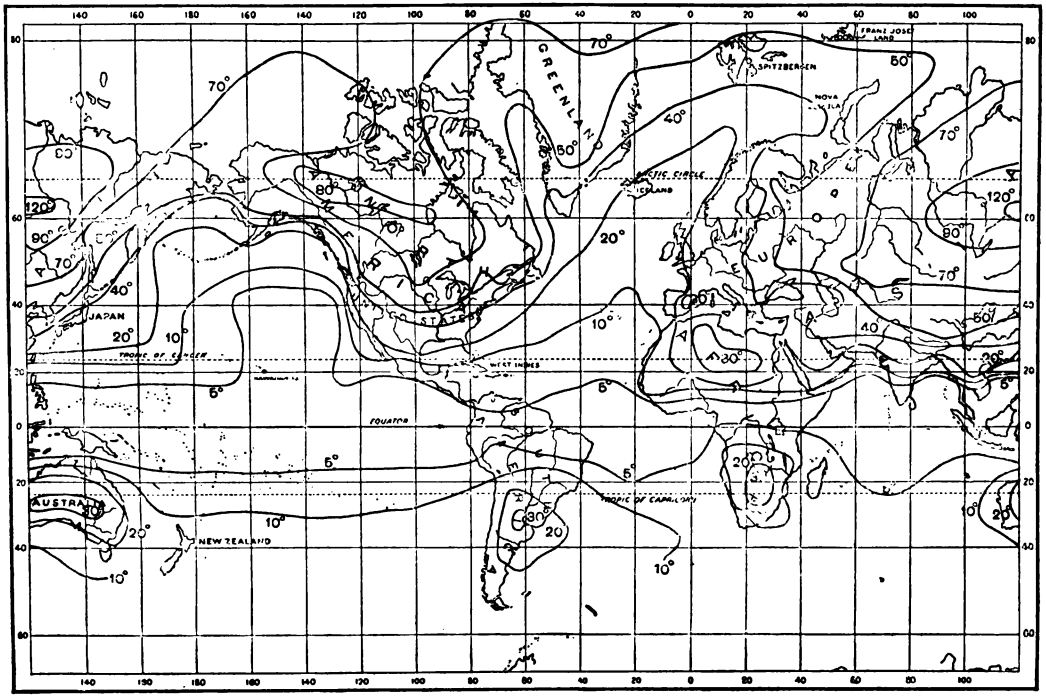

FIG. 14.—ANNUAL RANGE OF AVERAGE MONTHLY TEMPERATURE (AFTER CONOLLY).

perature. Such a chart is given in Fig. 14; and it shows that along the equator the annual range of temperature is 5° F.; and there is an increase with the distance from the equator, slowly over the oceans, but rapidly over the interiors of the continents.

In the southern hemisphere the temperature range reaches 30° F. in Africa, South America, and Australia. In the northern hemisphere the temperature range increases, at first slowly, and then at the extreme north rapidly, to about 40° F. in the Pacific and Atlantic oceans; but the increase is to 80° F. in north-central North America, and to 120° F. in northeastern Asia.

The 5° F. line, extending nearly across the South Pacific Ocean, shows a remarkable instance of uniformity in the temperature range over a water surface; and the 5° F. line in Africa, and the 40° F. and the 70° F. line in southern and central Asia, show for the interior of continents a like parallelism with the circles of latitude.

Charts of Average Highest and Lowest Temperatures for the Year. — The accompanying charts (Figs. 15 and 16) show the average maximum and average minimum air temperatures for the year over the whole earth. The lines of equal maximum and minimum temperatures are drawn for each 9° F. It must be understood that the charts represent the averages, for a number of years, of the extreme highest and lowest temperatures occurring in each year.

The average highest temperature for the year decreases with increase of latitude and altitude, but increases towards the interior of continents. The average lowest temperature for the year becomes lower with increase of latitude and towards the interior of continents.

Chart of Average Extreme Maximum Temperatures for the Year (Fig. 15). — This chart shows that the maximum temperatures have a quite regular distribution on the oceans, where, for wide zones extending both sides of the equator, there is a maximum of about 86° F.; and in no

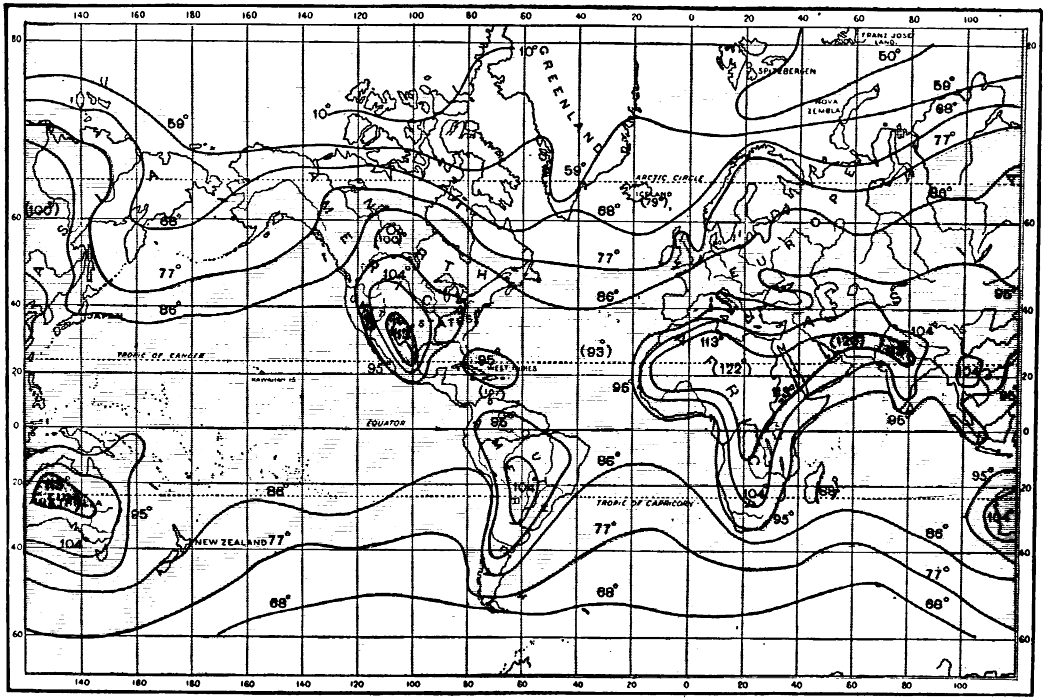

FIG. 15.—AVERAGE ABSOLUTE MAXIMUM TEMPERATURES (ABSOLUTE MAXIMUM IN PARENTHESES.

case does it reach 95° F. From these zones poleward there is a relatively rapid decrease with the latitude; and the temperature of 68° F. is reached near 60° latitude in the northern, and 50° latitude in the southern hemisphere.

The continental distribution is quite different, however, for there an increase of the maximum temperatures takes place with the progress inland. In the interior of northern Africa, Persia, northern India, Australia, and southern North America, temperatures of 113° F. are to be found; and in the southwestern United States, and perhaps in the Sahara desert, they even reach 122° F.

At high altitudes the extreme ranges of the maxima are less than on the lowlands, and are thus more like those for marine localities.

Chart of Average Extreme Minimum Temperatures for the Year.— This chart (Fig. 16) shows very strongly marked regional characteristics. On the Pacific, Atlantic, and Indian oceans, in the equatorial region, there are extensive zones stretching from west to east, in which the minimum temperatures do not go below 68° F. To the north of these there is a rapid decrease in the minimum temperatures, but to the south it is not so rapid.

There is rapid decrease towards the interior of the continents, and especially where there exist mountain ranges to cut off the access of the sea air to the interior. In the northern hemisphere there are three centers of extreme minimum temperatures,—one in the eastern part of Siberia, another in the northern part of North America, and a third in the interior of Greenland.

The 32° F. line is very interesting as showing the limits of the region in which the freezing point of water is reached. In the *northern hemisphere*, commencing at the Yellow Sea, we can follow its easterly course with a northern bend on the Pacific Ocean, cutting North America at about latitude 30°. On the Atlantic Ocean, following the Gulf Stream for a distance in a northeasterly direction, it reaches almost to Ireland, when it takes a southeasterly trend, and passes along the coast of the Spanish peninsula. Continuing through the Mediterranean Sea, it passes through southern Asia to the place from which we started in tracing it. In the *southern hemisphere* it makes almost a loop within the continent of Australia, and passes from its southeastern corner in an easterly direction through northern New Zealand, and thence onward until the southern end of South America is almost reached; but, stopping short of it, the line suddenly makes a northward curve up along

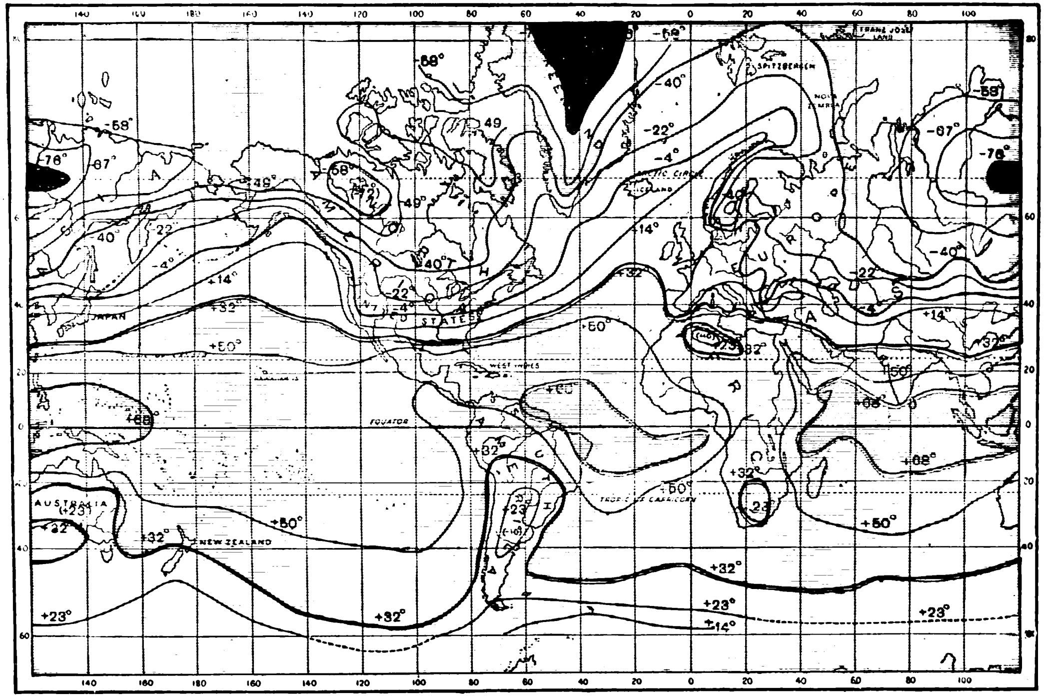

FIG. 16. — AVERAGE ABSOLUTE MINIMUM TEMPERATURES (ABSOLUTE MINIMUM IN PARENTHESES).

the coast to about latitude 20° south, where it crosses the continent, and then runs southward again along the eastern coast to latitude 45° south, and thence takes an easterly course to the southern end of Australia.

Variations in altitude above sea level have even a less influence on the minimum temperatures than on the maximum temperatures, as frequently in the severest cold weather there is an inversion of the temperatures with the altitude.

Average Annual Oscillation of Temperature. — The average extreme oscillation or amplitude of the temperature during the year, increases with the latitude and towards the interior of continents, but decreases with the altitude.

The lines of equal amounts of the extreme temperature oscillation during the year are drawn on the chart (Fig. 17) at intervals of 9° F.

These curves bring out most clearly the difference between the continental and marine climates. On the ocean near the equator the absolute temperature amplitude becomes less than 18° F. It increases, however, towards the poles and towards the continents, and especially towards the interior of the continents. In eastern Asia an average amplitude of 171° F. (in single cases over 180° F.), and in North America 153° F., is reached; but at the interior of Australia the amplitude is only 90° F., and in South America only 81° F. A further description of these charts cannot be given here, but they will well repay a careful study. In many respects they are of more interest than those of the average monthly or annual temperatures.

The average time of the lowest temperature for the year occurs earlier for a continental climate than for a marine climate. The warmest and coldest days of the year do not occur especially frequently on the average dates of extreme temperature, or on any other particular dates.

The Snow Line as Dependent on Temperature. — The *snow line*, or line of eternal snow, is that line along which, in the course of a year, just as much snow falls as can be

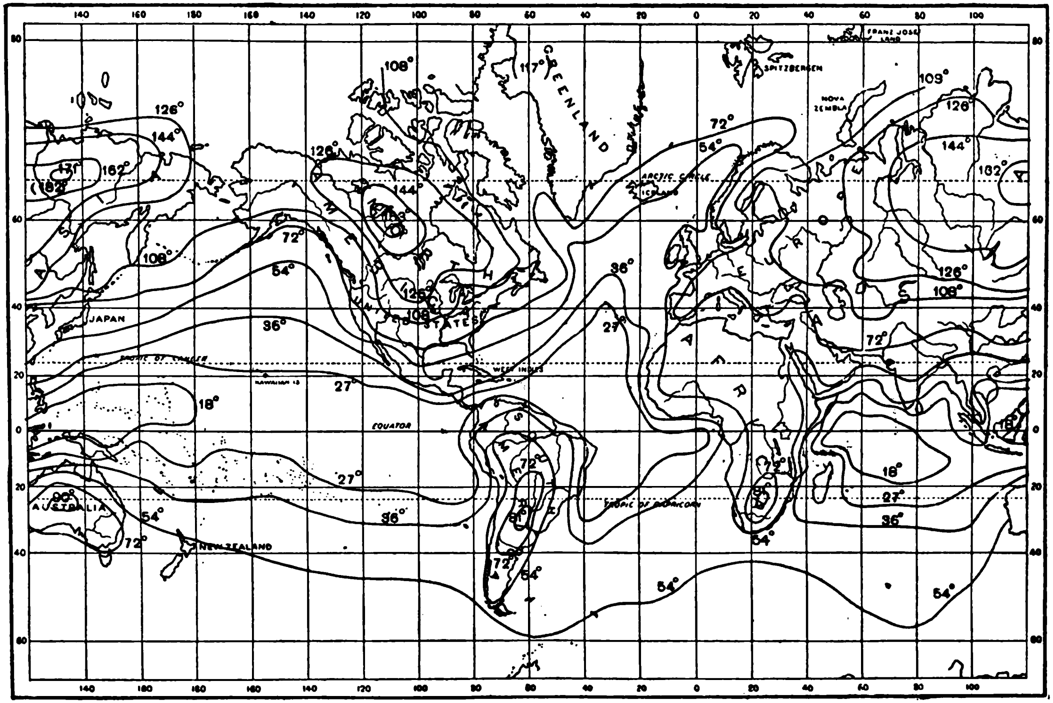

FIG. 17. — AVERAGE EXTREME TEMPERATURE OSCILLATION DURING THE YEAR (AFTER VAN BEBBER).
(ABSOLUTE OSCILLATION IN PARENTHESES.)

melted by the sun in the same time. Below this the snow would disappear; and above it, it would remain permanently present. This snow line is not fixed, but varies in altitude in different regions, and from year to year in the same region. The region of the snow line, then, is an altitudinal zone within which this line oscillates up and down with the varying temperature of the seasons.

The limits of this zone, and even the average height of this line, may be determined not only by means of many years' observations of the snow phenomenon direct, but also by means of the study of permanent glaciers. These last must extend up into the regions of eternal snow, and the upper limit of the snow line is lower than the peak surmounting the glacier. By an examination of the surrounding lower peaks, which are favorably situated for glacier formations, but do not have them, a lower limit for the glacier can be established; for the snow limit must be above these glacierless peaks, and below the peaks which have glaciers.

The temperature at the limit at which the snow disappears depends a good deal on the amount of snow which has accumulated. The mean annual temperature at the lower limit of the snow line varies from 37° to 3° F. for different regions for which we have observations; and it may be said, that, the greater the differences between the winter and summer temperatures, the less will be the average annual temperature at which snow is always found. Observations covering a number of years are necessary for determining the average position of the snow limit, which is more constant in lower than in higher latitudes.

The average annual temperature at which the snow line is found decreases towards the pole, as is seen from the table given on the next page. The snow line is found nearer the sea level as the cold polar regions are approached.

There has been a great deal of data, of more or less accuracy, accumulated concerning the snow limit in various parts of the world, but the following table will suffice for reproduction here: —

Mountains.	Geographical Latitude.	Altitude of Snow Limit above Sea Level.	Average Annual Temperature.
		Feet.	F.° about
Andes (near Quito) . .	Equator.	15,500	34
East African Mts. . . .	Near Equator.	15,500	
Himalaya (south side) .	27°–34° N.	16,000	31
Himalaya (north side) .	27°–34° N.	18,500	27
Middle and West Alps .	46° N.	9,000	27
Tyrolean Alps (central) .	47° N.	9,000 +	25
Nova Zembla	73° N.	2,000	12
Spitzbergen	77° N.	1,500	14

Temperature of the Ground. — The heat of the ground at or near the earth's surface is received from two sources, — the interior of the earth, and the sun. The heat of the earth increases towards the interior, and there is a flow of heat by conduction from the interior outward to supply the heat lost by radiation from the surface of the earth. The heat from the sun raises the temperature of the earth's surface higher than that due to the heat received only from within. According to the law for the conduction of heat, there must, then, be a transference of heat from the earth's surface towards the interior of the earth, and there is a decrease in the temperature of the successive layers of earth as far as the heat from the surface penetrates; and from this point inwards towards the center of the earth there is an increase of temperature.

The *diurnal changes of ground temperature*, following those of the air above, are greatest at the equator, and

decrease towards the pole, and are noticeable to a depth of about 40 inches at the equator. With descent below the surface, the times of maximum and minimum temperature are retarded.

The *annual changes of ground temperature* follow after those of the air above to a depth of about 80 feet; but the times of maximum and minimum temperature are retarded as much as several months at such depths.

The average ground temperature at a depth of about 3 feet below the surface is about 2° F. higher than the air temperature above. The normal increase in the earth temperature at greater depths is about 1° F. for each 60 feet descent; but very different rates of increase have been observed in various parts of the world.

The Region of Frozen Earth, where, at some depth below the earth's surface, there is a continuous temperature of 32° F. or a lower temperature, occurs in the polar regions and some very elevated mountains on other parts of the earth. The southern limit in the northern hemisphere at which the ground is perpetually frozen, is along the annual isotherm of about 28.5° F. for air temperatures at sea level.

Temperature of the Ocean. — Observations of the surface temperatures of the ocean water exhibit a daily change somewhat similar to that of the air above it; but the amplitude or range amounts to only a fraction of a degree during the 24 hours. The maximum temperature occurs shortly after midday, and the minimum just after sunrise.

The annual change of the ocean surface temperatures follows that of the air; but the amplitude is much less, and increases with the latitude up to high middle latitudes, where it decreases again. The times of maximum and

minimum temperatures are retarded about a month later than those for the air temperatures, and occur in August and February respectively, in the northern hemisphere.

In the warmest parts of the tropical zone the surface temperature changes very little during the year, and is about 82°–84° F. In middle latitudes it varies from about 50° F. in winter to 68° F. in summer. In the polar zone the temperature approaches or goes below the freezing point of fresh water, and varies but little with the season.

Temperatures of Small Isolated Bodies of Water. — The distribution of temperature in the lakes is regulated by the currents which arise from the alternate heating and cooling of the surface water. The warm surface water in summer imparts little of its heat to the water below, which consequently does not get much above 39° F. for deep lakes. There is, however, a condition of stable equilibrium, because the temperature decreases with the depth, and the coldest and heaviest water is at the bottom. In the fall and early winter the surface cools to about 39° F., at which temperature water is heaviest; and then, when the whole body of water reaches this temperature, the water is in indifferent equilibrium, and there are no vertical currents. When the surface water becomes still colder, it becomes lighter than the water below, and does not sink.

CHAPTER III.

AIR PRESSURE.

Nature of Air Pressure. — The atmospheric air, obeying the law of gases, exerts a pressure in all directions, and the amount of this pressure varies according to the density of the air. The air at the sea level, weighted down by the air above it, exerts a pressure of nearly 15 pounds per square inch of surface against which it presses.

The air pressure decreases with increase of altitude, because there is less air above, the higher the ascent. We do not know just how far from the earth's surface the air pressure is a measurable quantity, but at an altitude of a few miles it becomes very small. Where the free air has access below the earth's surface, there is a continued increase of air pressure with the descent. Also at various points on the earth's surface the air pressures are not quite the same; and for any one point the pressure undergoes variations of the nature of oscillations occurring in shorter or longer periods of time.

Distribution of Air Pressure. — The pressure of the air — called the barometric pressure, from the instrument used for measuring it — would be symmetrical, and would be practically the same at all places having a common altitude above the level of the sea, if it were not for the disturbing influence of the solar heat and of the movements

of the air caused by its unequal heating at different places on the earth.

If, under the condition of constant temperature, it is necessary to ascend a certain distance above the earth's surface in order to leave say $\frac{1}{10}$ of the atmosphere below and have the other $\frac{9}{10}$ above, then it is necessary to ascend the same distance in order to leave below $\frac{1}{10}$ of

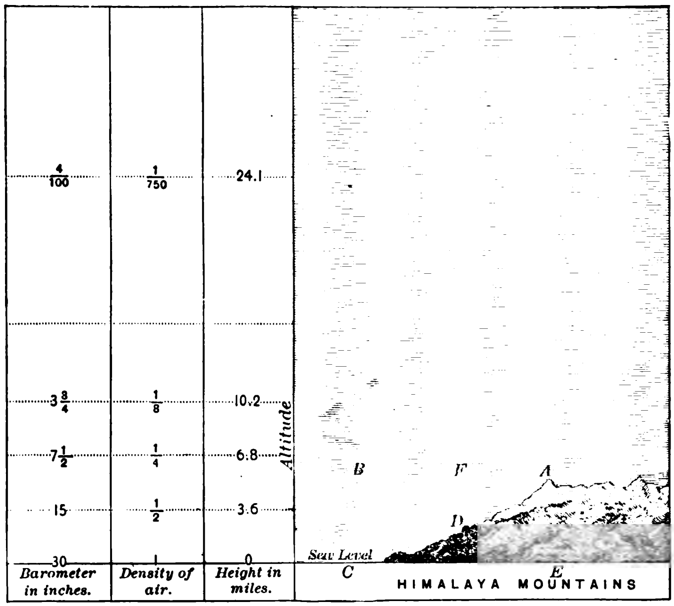

FIG. 18. — DECREASE OF DENSITY AND AMOUNT OF AIR WITH INCREASE OF ALTITUDE.

the remainder, or $\frac{1}{10}$ of $\frac{9}{10}$ of the whole (in addition to what was left below the first time); and so on up to any limit.

In the actual case of the atmosphere there is a decrease of temperature with the increase of altitude, and the air consequently becomes rarer at a progressively more rapid rate as the ascent is made; so that it is not necessary at higher elevations to go through so great a distance to get above a given proportion of the whole air as at lower elevations, where the air is warmer.

The accompanying diagram (Fig. 18) shows roughly the relative density of the air at various altitudes above sea level up to about the probable limit where the pressure of the air ceases to be a barometrically measurable quantity.

On the right of the diagram is shown the relation of the highest mountains to these altitudes. The column on the left, showing the pressure of the air at the various altitudes, will be better understood after reading the descriptions of barometers, the instruments by which such measurements are made. It is seen that the tops of the highest mountains have nearly three fourths of the total amount (by weight) of the air below them; and at the highest altitudes permanently inhabited by human beings about half the air is left below.

The accompanying table shows the relation of air pressure and altitude for every change of two inches in pressure from 30 to 16 inches. The altitudes are given in round numbers to the nearest 100 feet.

Altitude above Sea Level. FEET.	Barometric Pressure. INCHES.
0	30
1,800	28
3,800	26
5,900	24
8,200	22
10,600	20
13,200	18
16,000	16

The Barometer. — Galileo discovered, over two hundred and fifty years ago, that the air exerts its elastic pressure in all directions. Shortly afterwards (in 1643) Torricelli gave us an easy method of measuring this air pressure, or barometric pressure as we shall usually call it, when he invented the barometer.

The simplest form of barometer is obtained by filling

with mercury a glass tube of a length of, say, three feet, and one third of an inch in diameter, and closed at one end. Hold the closed end uppermost and insert the mouth of the tube in a cup also containing mercury (Fig. 19). On making the tube vertical, some of the mercury will flow downward out of the tube into the cup, until the weight of the mercury in the tube is counterpoised by the hydrostatic pressure of the air on the surface of mercury in the open cup.

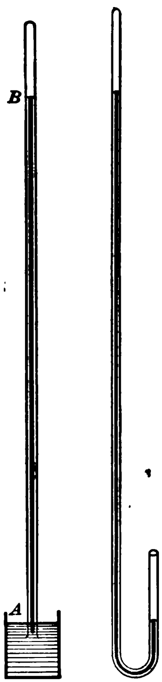

FIG. 19. – SIMPLE FORMS OF BAROMETERS.

The height of the surface of the mercury in the tube above the surface of the mercury in the cup is measured by means of a scale of length; and the amount, expressed in inches, is called the *barometer height*, or the *barometric pressure*, or merely the *barometer reading*.

In making accurate determinations of the barometric pressure, several sources of error must be allowed for. If proper care is taken in the manipulation of the tube, the space above the mercury in the barometer tube will be practically a vacuum, and consequently there will be no pressure on the top surface of the mercury in the tube. Fig. 19 shows on the left a barometer tube situated as just described. The figure on the right shows the position the mercury would take if the tube had a U bend at the bottom, in which case the mercury cup would be unnecessary. The distance *AB* is about 30 inches under ordinary air pressures at sea level.

Fig. 20 shows an actual barometer where the glass tube is incased in a metal tube which serves not only to protect it, but also for a measuring scale. The cup or cistern of mercury is at *A*. By means of the screw *C* the mercury is forced up until the mercury surface in *A* is brought

FIG. 20. — COMPLETE MERCURIAL BAROMETER.

to the little pointer shown at *B*, which is the zero or beginning of the barometer scale. The height of the mercury in the tube is then read off from the scale at *C*. *D* is an attached thermometer.

Fig. 21 shows an aneroid barometer. This is an instrument shaped something like a small clock, and the barometric pressure is read by means of a hand or indicator, on a scale of inches placed on the face like the minute divisions of a clock. A little flat, circular, metal box within the case is partly exhausted of air, and tightly sealed; and when the outer air pressure increases, it forces in the side of the box, and this motion is communicated to the hand, making it move around toward the right. When the pressure is lessened, the hand moves to the left. Another hand, which is usually placed on the case, remains stationary unless moved by some person.

FIG. 21. — ANEROID BAROMETER.

The Reduction of Barometric Pressure to Sea Level. — This is one of the most unsatisfactory problems connected with practical meteorology. The observations of air pressure are usually made at places having various (known) altitudes; and, since the air pressure decreases with altitude, then, in order to compare the results from these places, the pressures must be reduced to some common plane having a fixed altitude. For convenience the sea

level has been chosen as a level for reduction, but the observations could be reduced to any other desired level, and sometimes other levels are chosen.

Closely connected with this matter is the determination of the difference in the altitude of two places by means of the observations of the air pressure at both places. We can suppose observations to be made on various parts of the mountains shown in Fig. 18. If the barometric pressure is observed in a balloon at *B*, and we wish to reduce this to the corresponding pressure at sea level, we should add it to the weight of the air column *BC* (expressed in barometric pressure); and similarly for reducing barometric pressure observed at *A* to sea level at *E*, we must add to it the weight of an assumed column of air, *AE*.

The problem of the reduction of the barometric pressure to sea level is too complicated to explain in full here, but it amounts to this, — that we have the air pressure given for a certain altitude, and it is required to find the weight of a column of air which would extend from this altitude down to sea level. In the case of an observation in a balloon over the ocean, there would be a real column of air to sea level; but in the case of an observation on a mid-continental mountain, the column of air would be purely fictitious. In order to determine the weight of the intervening column of air, it is necessary to know its average temperature. This might be determined in the case of the balloon; but for the mountain the local temperature of the region must be taken as the upper temperature, and from this the temperature at the assumed sea level below must be calculated, in order to obtain the average temperature of the fictitious air column, so that its weight can be computed. The laws of the decrease of air temperature with altitude are so different for different regions, that there is room for considerable error in obtaining the average temperature of the air column by this means; and in proportion as this is in error, just so much is the computed weight of the air column in error. When the weight of the air column is determined on the barometric scale, this amount is added to the corrected observed barometric pressure, and the sum is the barometric pressure reduced to sea level. Tables for facilitating the

reduction of barometric observations to sea level are numerous. Those in the "Smithsonian Meteorological Tables" and Hazen's "Meteorological Tables" are very complete.

The Determination of the Difference in Altitude of two neighboring places by means of simultaneous barometer observations at the two points, is the reverse of this reduction from one known level to another known level. In this case the barometric pressure and air temperature are given for the two points; and it is required to determine the vertical length of a column of air which has the pressure and temperature of the higher place at its top, and those of the lower place at its base. If we wish to determine the difference in altitude between *A* and *D* (Fig. 18), we should observe simultaneously the barometric pressure at *D* and at *A* (which would be the same as at *F*), and the difference in the pressures will be the weight (in pressure) of the column of air *FD*.

We can determine the vertical length of the air column for each inch of difference in these barometric pressures, and, by combining their whole and fractional parts, obtain the total distance corresponding to the whole difference of pressures.

Tables are also used to facilitate these computations; and since the difference in the altitudes of the two places seldom extends from high elevations to the sea level, and since the temperatures of the lower station are known, these reductions and computations are usually capable of being carried out with greater accuracy than the reductions to the sea level, which have been described just above.

As it may be useful to make a rough computation of the reduction of barometric pressure to sea level when the altitude, and consequently the vertical length, of the air column, is known, or to determine the difference in altitude when the approximate height of the column of air (equal to a difference in barometer readings) is given, the following table is offered for such use. It gives the vertical length, in feet, of a column of air corresponding to 0.1 of an inch of the barometric pressure at various air pressures and temperatures.

VERTICAL LENGTH OF AN AIR COLUMN CORRESPONDING TO 0.1 OF AN INCH BAROMETRIC PRESSURE.

AIR PRESSURE IN INCHES.	AVERAGE TEMPERATURE IN DEGREES FAHRENHEIT.						
	20°	30°	40°	50°	60°	70°	80°.
	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.
22	116	119	122	124	127	130	132
23	111	114	116	119	124	124	126
24	106	109	111	114	116	121	121
25	102	105	107	109	112	114	116
26	98	101	103	105	107	110	112
27	94	97	99	101	103	106	108
28	91	93	95	98	100	102	104
29	88	90	92	94	96	98	100
30	85	87	89	91	93	95	97

Thus, when the barometer reads 28 inches and the temperature is 50° F., then at a point 98 feet lower in altitude the barometer will read 0.1 of an inch more than 28 inches, or 28.1 inches; or at a point 98 feet higher it will read 27.9 inches.

Results of Observations of Atmospheric Pressure. — The observations of barometric air pressure are not so frequent as those of air temperature, on account of the cost, complexity, and fragile construction of barometers. Still, enough observations have been made to allow the conditions and changes of the atmospheric pressure to be noted in most of the accessible parts of the surface of the globe. The observed conditions of atmospheric pressure have been studied locally, that is, for individual places; and also in their geographical distribution, that is, in their connection with those of other places.

Where but a single place is concerned, it is usual to consider the observed pressures, correcting them for instrumental errors only; but where the geographical distribution is to be considered, it is necessary to reduce the observations to a common level.

Observed barometric pressures at individual places show the existence of both diurnal and annual periodic changes, somewhat similar to those found for temperatures, but in a less degree. These periodic changes are, however, masked by the far greater accidental changes which must be eliminated (by taking the average of many observations) before the periodic changes are visible.

The Diurnal Change or March of the Air Pressure does not culminate, like the temperature, in a single maximum with a corresponding minimum, but there are two maxima and two minima. In general, the minima occur at about 4 h. and 16 h. (4 A.M. and 4 P.M.), and the maxima at about 10 h. and 22 h. (10 A.M. and 10 P.M.).

The *amplitudes of oscillation* vary in different parts of the world, but are always small, amounting to only about 0.15 of an inch in the regions of greatest, and to 0.01 of an inch in the regions of least oscillation. The amplitude is in general greatest near the equatorial regions, and diminishes towards the poles.

The hourly pressures for a few stations are given in the following table: —

TABLE SHOWING DAILY MARCH OF THE BAROMETRIC PRESSURE ON THE AVERAGE FOR THE YEAR (*hourly barometric pressures in inches and hundredths*).

	1^h	2^h	3^h	4^h	5^h	6^h	7^h	8^h	9^h	10^h	11^h	12^h	13^h
Key West, Fla. .	30.07	.06	.05	.05	.05	.06	.07	.08	.09	.10	.09	.08	30.07
St. Paul, Minn. .	30.13	.13	.13	.13	.14	.14	.14	.15	.15	.15	.15	.14	30.12
Fort Conger . .	29.844	.846	.847	.849	.849	.848	.847	.846	.844	.842	.839	.838	29.839

	14^h	15^h	16^h	17^h	18^h	19^h	20^h	21^h	22^h	23^h	24^h	Av.
Key West, Fla. .	30.05	.04	.03	.03	.04	.05	.06	.07	.08	.08	30.08	30.06
St. Paul, Minn. .	30.11	.11	.11	.11	.11	.11	.12	.12	.13	.13	30.13	30.13
Fort Conger . .	29.841	.844	.846	.846	.846	.845	.845	.845	.844	.842	29.843	29.844

It is to be noticed that for the Arctic regions it is necessary to give the pressures in thousandths of an inch in order to show the diurnal change.

In the accompanying figure (Fig. 22) the daily march of the barometric pressure is given for Key West, St. Paul, Fort Conger in the Arctic region, and for Calcutta, India. In this diagram the average daily barometric pressure is entered as the horizontal line marked 0.00 inches; and the departure or the amount above or below this at the different hours is shown on the different lines, and the rise and fall by the course of the curves.

The *principal maximum* barometric pressure occurs before noon, and the *principal minimum* after noon. During the winter season they approach noontime, but in summer they recede from it. The amplitudes of the diurnal oscillations are greater during the warmer than during the cooler part of the day, and for the dry than for the moist continental localities. They diminish from the equatorial region of greatest solar heat towards the poles. The cause of these oscillations is not at present perfectly understood; but they may be due to waves in the atmosphere which depend in some way on the intensity and distribution of the solar heat.

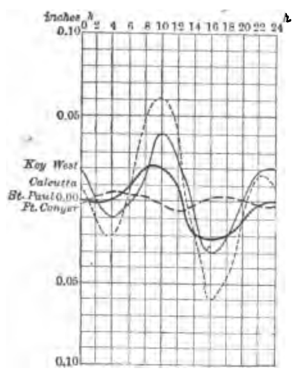

FIG. 22. — DIURNAL CHANGES OF AIR PRESSURE.

The Annual Change or March of the Air Pressure, that is, the change in the average pressure from month to month, presents a variety of characteristics for different regions of the earth. On the continents at the lower altitudes there is a maximum air pressure in winter, and a minimum in summer. On the ocean, on the contrary, the air pressure is highest in summer, and lowest in winter; and this is also true of the air pressure at high altitudes.

This matter cannot be fully understood until the motions of the atmosphere are studied; but it may be said in general, that over the continents in the summer the air becomes very much heated, and the surfaces of equal air pressure more elevated, which causes the air to flow outward (above) towards the oceans, while the air must flow inward (below) from the oceans towards the continents to compensate for the outflow above. Thus the warmer continental air causes a deficiency in the amount of air, and consequently a diminished air pressure, over the continent.

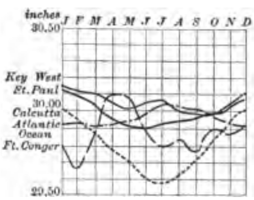

FIG. 23. — ANNUAL CHANGES OF AIR PRESSURE.

The greatest and most regularly occurring variations in the mean air pressure take place on the continents, and the least on the oceans and their coasts. There are, however, many local deviations from this law.

In the interior of Asia the amplitude of oscillations amounts to over 0.70 of an inch in a year; while over the Atlantic Ocean near the equator it is only about 0.10 of an inch.

The following table shows the average monthly air pressures at a few points on the earth's surface: —

ANNUAL MARCH OF AIR PRESSURE (*monthly barometer pressure in inches and hundredths*).

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
Calcutta . . .	30.01	29.95	29.86	29.76	29.70	29.58	29.57	29.62	29.71	29.84	29.95	30.01	29.80
Key West . . .	30.14	30.12	30.10	30.04	30.01	30.04	30.06	30.01	29.99	29.99	30.06	30.12	30.05
St. Paul	30.13	30.10	30.04	29.95	29.91	29 90	29.93	29.95	29.97	29.99	30.04	30.09	29.99
Fort Conger . .	29.80	29.67	29.89	30.10	30.07	29.88	29.79	29.83	29.77	29.90	29.86	29.92	29.88
Atlantic Ocean .	29.93	29.94	29.92	29.93	29.94	29.98	30 04	30.02	30.01	29.97	29.94	29.92	29.96

The data in this table are shown graphically in Fig. 23.

The Irregular or Non-periodic Oscillations in the Air Pressure are those changes which occur more or less

gradually from time to time, and which do not belong to the regular diurnal or annual periodic changes. They are due to the movement, over the earth's surface, of areas of abnormally high or low barometric pressure.

The *monthly barometer oscillations* are the difference between the highest and lowest barometer reading during the month. The following table shows the average barometric oscillations in the northern hemispheres, over the sea and over the land, for the winter and the summer seasons.

AVERAGE MONTHLY BAROMETRIC OSCILLATIONS (*inches and hundredths*).

NORTH LATITUDE.	IN WINTER.		IN SUMMER.	
	OVER THE OCEAN.	OVER THE CONTINENTS.	OVER THE OCEAN.	OVER THE CONTINENTS.
0°	0.20	0.24	0.20	0.24
10°	0.24	0.32	0.20	0.24
20°	0.32	0.43	0.24	0.32
30°	0.63	0.51	0.35	0.43
40°	1.14	0.71	0.63	0.47
50°	1.50	0.98	0.98	0.55
60°	1.77	1.22	1.10	0.75
70°	1.57	1.14	0.98	0.71
80°	1.34	—	0.71	—

The place of greatest oscillation is in the North Atlantic Ocean, between Newfoundland and the British Islands, where it amounts to nearly 2 inches in winter. From this locality there is a decrease in the amplitude of oscillation, quite gradual on the north, east, and west sides, but very rapid towards the south; and in the neighborhood of the equator it becomes almost identical with the periodic daily oscillations.

These barometer oscillations are much greater over the sea than over the land; and greater in winter than in summer, except near the equator, where they are quite uniform, as may be seen from the table just given.

Graphical Representations of the Distribution of Air Pressures. — An *isobaric line*, or an *isobar* as it is usually called by meteorologists, is a line drawn through adjacent places which have the same barometric pressure. In order to draw isobaric lines on a chart, the barometric pressures reduced to sea level are written down on the map at the places where the observations are made. Then the isobaric lines are drawn on the map usually for each even tenth of an inch of barometric pressure.

An *isobaric surface* is a surface in the air all points of which have the same barometric pressure.

When we wish to show the position or distribution of isobaric lines and isobaric surfaces in different parts of the atmosphere by means of diagrams, it is usual to take some plane or level surface, such as the sea level or some imaginary surface parallel to it, and designate where the isobaric surfaces intersect this plane, and the positions which they assume with reference to it. The lines where the isobaric surfaces cut this plane are isobaric lines, or isobars.

There are two ways in which the level of isobaric surfaces is disturbed: —

1. Air may be added to a region or taken from it by means of air currents. In case air is added, the pressure is increased, and the isobaric surfaces are forced downward towards the earth's surface, and lie closer together than in adjacent regions where the amount of air is diminished or unaltered. In case air is abstracted, the pressure is diminished, and the isobaric surfaces are forced upwards, and lie farther apart than in adjacent regions. Such changes in the isobaric surfaces are made manifest all the way down to the earth's surface, and at the ground there is an increase or decrease of the air pressure.

2. The quantity of air over any given region may remain unchanged, but the temperature may be varied; in which case there is no increase or diminution of the air pressure at the ground, and the isobaric surfaces there remain unchanged, but at various elevations above the ground the positions of the isobaric surfaces will depend on the distribution of the temperature.

This last condition has been chosen for showing in more detail the nature of isobaric surfaces and isobaric lines.

Let us suppose that we are up in the air at some distance above the ground, and that we take the plane or level of this page held horizontally as a chosen level in the air, having a barometer reading of 29 inches, for example. It becomes necessary, then, to imagine the leaf of the book to be so perforated that the air can pass freely through it, and that no resistance is offered by it to the air moving either upwards or downwards.

Suppose that the temperatures are uniform on this level, and there are also common temperatures at other levels above and below. Then for this region, inclosed by the circle in Fig. 24 (1), the barometric pressures are 29 inches. Now, let the temperature of the air become higher in the center at *A*, and for some distance above and below the plane, and colder on the other edge at *B*, but so that the present condition of temperature and pressure is maintained at the points on a circle, *C*, surrounding *A*, and between *A* and *B*, Fig. 24 (2). Then, since warmer air is expanded and colder air is contracted, the warmer air at *A* on the level of the page will in Fig. 24 (2) be forced upwards above the page towards the reader's eye, while the colder air at *B* will be contracted below or beneath the page; but at *C* it will remain on the level of the page. When more air is forced above the plane of the page at the center *A*, the former pressure of 29 inches on the page is increased by the amount of air pushed up; while on the outside, *BB*, the air pressure on the page is diminished by the amount of air drawn below the level of the page. But around the circle *C*, where the temperature remains the same, the air pressure remains 29 inches.

We will suppose that the pressure on the level of the page at *A* is increased to 29.5 inches, and that on the outskirts *BB* it is reduced to 28.5 inches, then we shall have the pressures on the level or plane of the page at *A*, *B*, and *C*, as shown in Fig. 24 (3).

The pressure on the level of the page will then be greatest at the center A, and will gradually decrease to the outer limits B. We can then also draw on the page the locations where any of the air pressures between 29.5 inches and 28.5 inches will occur by taking the proportional distances. The air pressure 29.4 will occur on the circle drawn around A at $\frac{1}{5}$ of the distance towards C; the air pressure

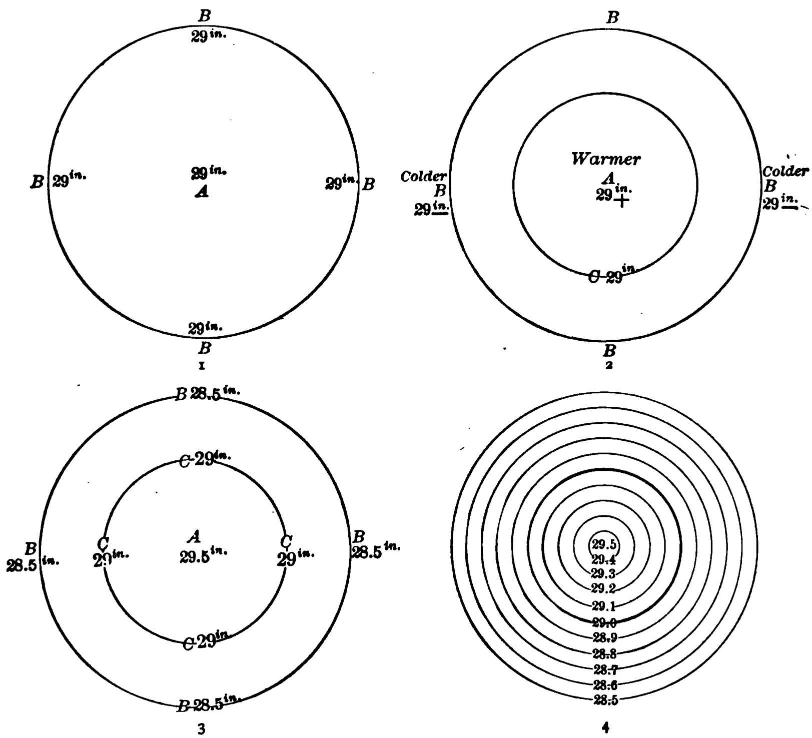

FIG. 24. — ILLUSTRATION OF LOCAL INCREASE AND DECREASE OF AIR PRESSURE.

29.3, at another $\frac{1}{5}$ of this distance, etc.: so that, if we wish to show the location of the pressures for each $\frac{1}{10}$ of an inch, we should have a diagram like Fig. 24 (4).

The numerals on these circles show the air pressure on this plane; and they, with the position of the circles, show us the relative loca-

tions, above and below this plane, of the isobaric surfaces which intersect it at these circles; and this is the best graphical presentation that we can give of them, because it is so difficult to represent, on a sheet of paper having only length and breadth, the additional feature of thickness.

It is by a similar graphical process that the air pressures over the earth are shown on charts. We trace out by isobaric lines the intersection of the isobaric surfaces with some chosen level or plane, that of the sea level being usually selected; and by so locating these isobaric surfaces as shown by observations, we can determine the regions of greatest and least air pressure.

Geographical Distribution of Air Pressure. — In comparing the average air pressures at various places, the observed pressures, as we have seen, must be reduced to a common level. It is usual to make the reduction to the sea level; but any level may be chosen, and, as we shall see, other levels are also sometimes used as a plane of reduction.

Charts have been prepared, showing the distribution, over the surface of the earth, of the average air pressure (reduced to sea level) for each month of the year. The isobaric lines or isobars drawn on these charts show very clearly the regions of relatively high and low air pressures, and the intermediate conditions.

Areas of relatively *high* air pressure are called *barometric maxima*, or sometimes simply *highs;* and areas of relatively *low* pressure, *barometric minima*, or simply *lows*.

Only selected charts are given here for the year, and for the months of January and July, on which are indicated the isobaric lines which show where the isobaric surfaces cut the plane of the sea level. The order of the distribution of the air or barometric pressure depends on two great causes, — primarily on the air temperatures, as shown by their distribution over the surface of the earth; and secondarily on the movement of the air itself from one part of the earth's surface to another. The discussion of the relation of these causes to their

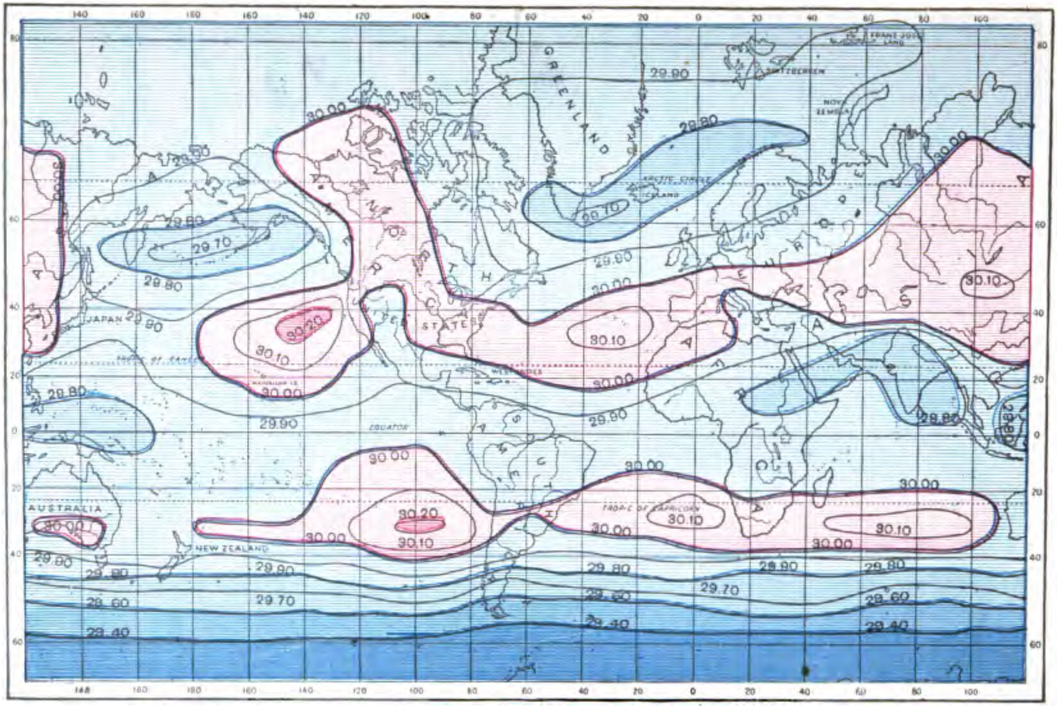

FIG. 25.—ISOBARS (AT SEA LEVEL) FOR THE YEAR (AFTER BUCHAN).

effects must be deferred until the movements of the atmosphere are presented in a later chapter. The mere facts of the distribution of the air pressure are all that can be given at present.

Distribution of the Average Air Pressure for the Year. — This chart of isobaric lines at sea level for the year (Fig. 25) shows, in the northern hemisphere, an area of *high* air pressure of more than 30.1 inches over eastern Asia; another with a pressure of more than 30.2 inches over the eastern North Pacific Ocean, in the same latitude; and another of more than 30 inches over North America. In the southern hemisphere the regions of high pressure are, over the Atlantic Ocean, near latitude 25°, more than 30.1 inches; over the Indian Ocean, west of Australia, more than 30.1 inches; over the eastern Pacific Ocean, between latitude 30° and 40°, more than 30.2 inches.

Areas of *low* air pressure are found, for the northern hemisphere, over the North Atlantic Ocean, near Iceland, with pressure below 29.7 inches; over the extreme northern Pacific Ocean, with pressure below 29.7 inches; and to the southeast and southwest and over the southern part of Asia, with pressure below 29.8 inches. In the southern hemisphere the regions of low air pressure are over northern Australia and over the Antarctic Ocean, where there is a quite uniform decrease from 30 inches just north of latitude 40°, to 29.3 inches in latitude 60°.

Distribution of the Average Air Pressure for January. — The chart of the isobaric lines at sea level, for the month of January (Fig. 26), shows that in the northern hemisphere there is an area of high barometric pressure over central Asia of more than 30.5 inches, and over central North America of more than 30.2 inches; in both cases lying mainly north of the 40th parallel. In the southern hemisphere the areas of maximum pressure of

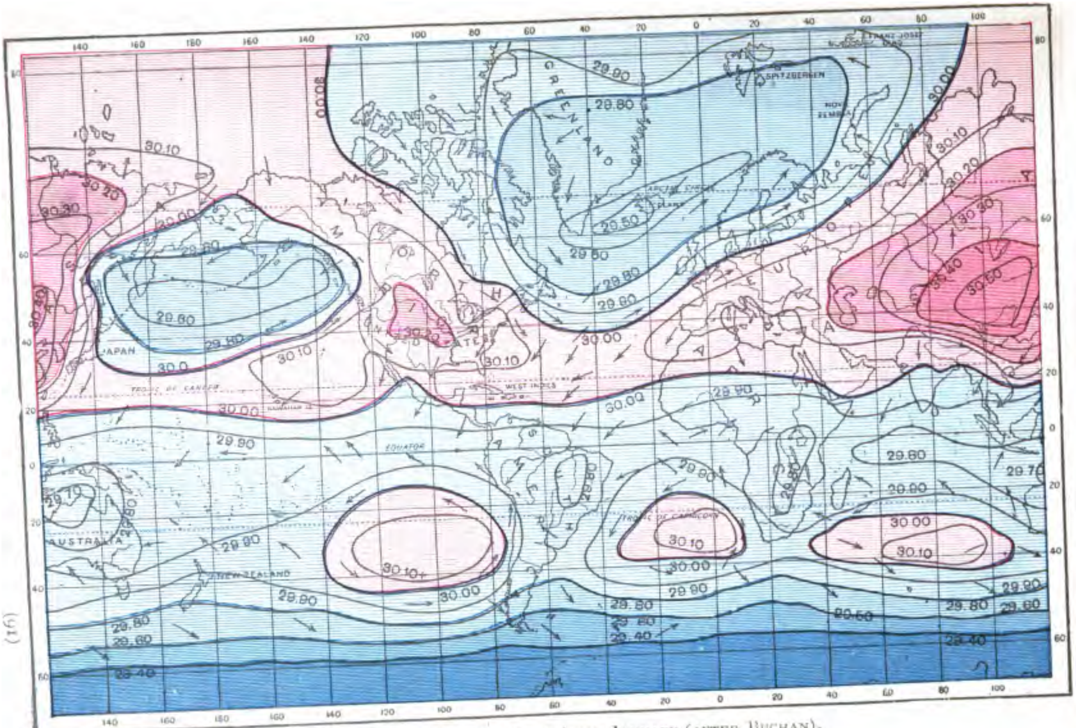

FIG. 26. — ISOBARS (AT SEA LEVEL) FOR JANUARY (AFTER BUCHAN).

30.1 inches are on the equatorial side of the 40th parallel over the oceans to the westward of Australia, Africa, and South America.

Areas of lowest pressure in the northern hemisphere are, over the North Atlantic, near Iceland, less than 29.5 inches; and over the central North Pacific, less than 29.6 inches. In the southern hemisphere the areas of lowest pressure are, over northern Australia, less than 29.7 inches; the Indian Ocean, less than 29.8 inches; and southern Africa, 29.8 inches. South of the 40th parallel there is a rapid but regular decrease from 29.9 inches, to about 29 inches in the Antarctic latitudes.

The distribution of the average air pressure for January reduced to planes above the sea level (as indicated by isobaric lines drawn for other elevations than that of the sea level as just given) shows that for the January average air pressures the above-mentioned maximum at the earth's surface over Asia exists only for the lower air layers; and at an altitude of about 5,000 feet, the North American and European maximum air pressures disappear. At an altitude of about 10,000 feet, the maximum air pressure lies over the equatorial region, and it decreases thence towards the poles.

The distribution of air pressure for January, at an altitude of 13,000 feet, is about as follows: —

Over the continents in the northern hemisphere there is a region of low pressure over northern North America (16.8 inches) and Asia (16.5 inches); but on the intervening oceans the pressure is higher, especially over the North Atlantic Ocean (17.3 inches). From these northern regions there is an increase with southward progress, more rapid over the land than over the sea, down to the region between the equator and the Tropic of Capricorn, where a maximum pressure of about 18.7 inches is reached. Within this maximum region the pressures vary somewhat, and are a little less over the Atlantic Ocean than elsewhere. From this region of high pressure there is a decrease towards the south, and a pressure of about 17.7 inches is reached at latitude 50° south.

It is to be particularly noticed that the courses of the isobars are

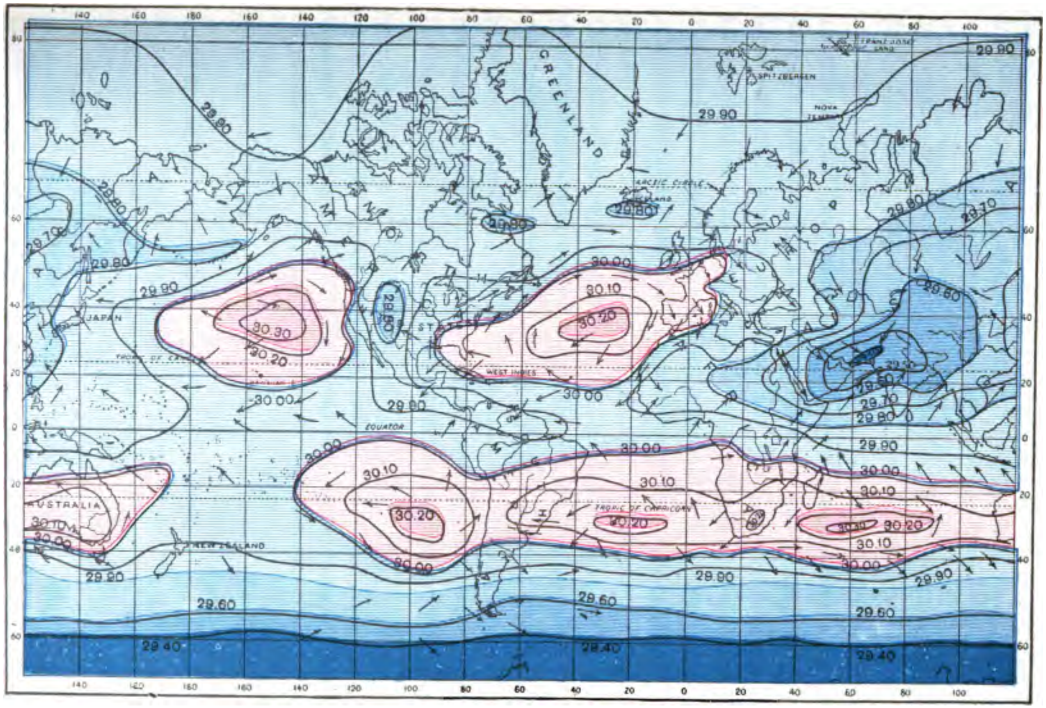

FIG. 27. — ISOBARS (AT SEA LEVEL) FOR JULY (AFTER BUCHAN).

30.1 inches are on the equatorial side of the 40th parallel over the oceans to the westward of Australia, Africa, and South America.

Areas of lowest pressure in the northern hemisphere are, over the North Atlantic, near Iceland, less than 29.5 inches; and over the central North Pacific, less than 29.6 inches. In the southern hemisphere the areas of lowest pressure are, over northern Australia, less than 29.7 inches; the Indian Ocean, less than 29.8 inches; and southern Africa, 29.8 inches. South of the 40th parallel there is a rapid but regular decrease from 29.9 inches, to about 29 inches in the Antarctic latitudes.

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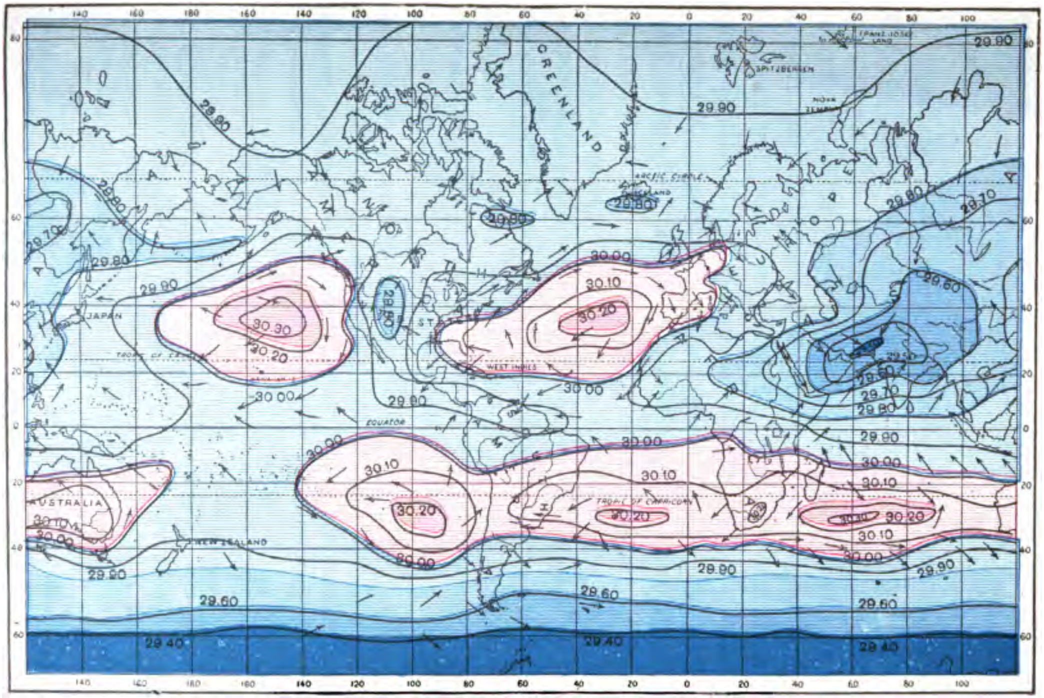

FIG. 27. — ISOBARS (AT SEA LEVEL) FOR JULY (AFTER BUCHAN).

very irregular over the northern hemisphere, where the earth's surface is mainly land, and that they run regularly along the parallels of latitude in the southern hemisphere, where the surface is mostly water.

Distribution of Air Pressure at Sea Level for July (Fig. 27). — The areas of high pressure for July in the northern hemisphere lie north of the Tropic of Cancer (over the Atlantic Ocean, 30.2 inches; and over the Pacific Ocean, 30.3 inches). In the southern hemisphere the areas of high pressure lie over the oceans, south of the Tropic of Capricorn (in the eastern Pacific and mid-Atlantic, 30.2 inches; and in the western Indian Ocean, 30.3 inches).

The areas of low pressure in the northern hemisphere lie over the continents, a little north of the Tropic of Cancer (in western Asia, 29.4 inches; in western America, 29.8 inches). There is also an area of low pressure over the Atlantic near Iceland (29.8 inches). In the southern hemisphere, south of latitude 40°, the pressure is below 30 inches, and decreases rapidly towards the Antarctic regions, being 29.4 inches in latitude 60°.

For July, at the higher planes above the sea level, the distribution of the air pressure differs from that just described. At about 5,000 feet altitude, the curves are much more regular than at sea level. At about 10,000 feet altitude, the zone of the highest pressure approaches the equator, and at 13,000 feet it reaches the equatorial region.

The distribution of air pressure for July, at an altitude of 13,000 feet, is as follows: —

The region of highest pressure lies mainly on the north side of the equator, and extends even beyond the Tropic of Cancer. This region of maximum pressure is somewhat broken; and over the center of Africa, southern North America, the Middle Atlantic and Pacific oceans, the pressure reaches maximum values of 18.6 inches or 18.7 inches, which is slightly above that of the intervening regions along the Tropic of Cancer. To the northward the pressure decreases to 17.8 inches over northern North America, and 17.9 inches over the North Atlantic and northern Asia; but the decrease is more gradual

over the western coasts than over the eastern coasts. To the southward of the equator the decrease is much more regular than towards the north, and at about latitude 55° south the pressure (17.6 inches) is without many irregularities with change of longitude; which shows, that, even at this altitude, the difference between land and water surfaces is noticeable.

Average Air Pressure along a Meridian. — The average change of air pressure with latitude, or the change along an average meridian, may be seen in the table on p. 96, which gives the average or normal pressure for each 5° of latitude for the year, January, and July, at sea level; and for the year at altitudes of about 6,500 feet and 13,000 feet.

For the year, at the sea level, there is a maximum air pressure at about latitude 35° in the northern hemisphere, and 30° in the southern; with a decrease on both sides towards the equator and towards the poles. This poleward decrease is very much greater in the southern hemisphere than in the northern, due to the effects of the uniform water surface in the former; and in fact in the northern hemisphere there is a slight increase again at the far north. The decrease toward the equator is very slight.

At 6,500 feet altitude above the sea level, the place of maximum pressure in each hemisphere approaches the equator, and is quite uniform near it, and the decrease towards the poles is much more nearly equal in the two hemispheres.

At 13,000 feet altitude, a single maximum occurs near the equator (a little to the south of it), and the diminution of the pressure towards the poles is still more symmetrical for the two hemispheres. The area of high pressure at the north pole at sea level seems to disappear at a slight altitude, just as it does in the more local region of eastern Siberia.

AVERAGE AIR PRESSURES ALONG A MERIDIAN.

LATITUDE.		Average Air Pressure at Sea Level.			Average Air Pressure for the Year at Altitude of	
		YEAR.	JAN.	JULY.	6,562 FT.	13,123 FT.
		Inches.	Inches.	Inches.	Inches.	Inches.
North	80° . . .	29.941	29.937	29.945	22.913	17.528
	75° . . .	29.921	29.929	29.874		
	70° . . .	29.866	29.882	29.850	22.976	17.583
	65° . . .	29.850	29.874	29.827		
	60° . . .	29.870	29.910	29.831	23.134	17.791
	55° . . .	29.910	29.961	29.858		
	50° . . .	29.949	30.004	29.894	23.347	17.992
	45° . . .	29.980	30.039	29.921		
	40° . . .	30.000	30.063	29.937	23.543	18.252
	35° . . .	30.016	30.083	29.949		
	30° . . .	29.988	30.055	29.921	23.658	18.437
	25° . . .	29.937	30.000	29.874		
	20° . . .	29.890	29.945	29.835	23.658	18.500
	15° . . .	29.854	29.894	29.815		
	10° . . .	29.839	29.858	29.819	23.658	18.532
North	5° . . .	29.843	29.843	29.839		
Equator	0° . . .	29.843	29.819	29.866	23.665	18.543
South	5° . . .	29.854	29.807	29.902		
	10° . . .	29.886	29.819	29.953	23.685	18.547
	15° . . .	29.929	29.850	30.008		
	20° . . .	29.988	29.902	30.075	23.728	18.547
	25° . . .	30.047	29.953	30.154		
	30° . . .	30.059	29.973	30.146	23.709	18.476
	35° . . .	30.016	29.945	30.087		
	40° . . .	29.941	29.886	29.996	23.508	18.232
	45° . . .	29.815	29.776	29.854		
	50° . . .	29.654	29.634	29.673	23.150	17.862
	55° . . .	29.457	29.457	29.457		
	60° . . .	29.268	—	—	22.717	17.476
	65° . . .	29.122	—	—	—	—
South	70° . . .	29.055	—	—	—	—

The accompanying figure (Fig. 28) shows the general slope of the isobaric surfaces along a meridian, at sea level and at various altitudes above it. The dotted line shows where the isobaric surface of 30 inches lies below the sea level, the pressure at that level being less than 30 inches.

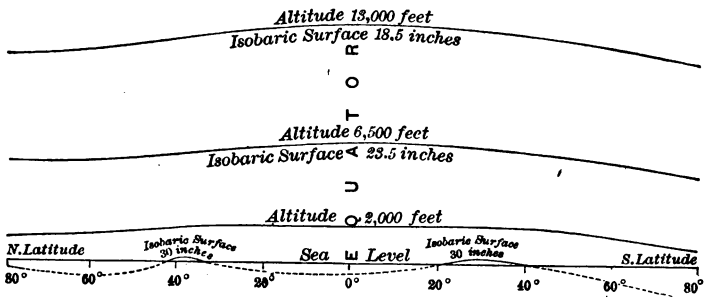

FIG. 28.—SLOPE OF ISOBARIC SURFACES ALONG A MERIDIAN AT VARIOUS ALTITUDES.

The Variability of the Average Air Pressure increases with the latitude, and to a certain extent with the increase of the distance inland. With increase of altitude above the sea level, the variability decreases in winter, but increases in the summer for low altitudes. The variability is greater in winter than in summer.

Long-period Oscillations of Air Pressure, extending over a period of a little more than 30 years, undoubtedly exist; but the extreme average maximum pressure for a number of places is not more than 0.06 of an inch greater than the extreme average minimum. Slight as these variations are, they are sufficient to cause great climatic oscillations, owing to their effects on the atmospheric circulation. It is very probable that when there is an excess of air pressure over the continents, there is a deficiency over the oceans.

Physiological Effects of Air Pressure.—The outside surface of the body exposed to the air pressure, in the case of a grown person, is about 16 square feet, and this would

give a pressure of about 35,000 pounds. If it were not for the ease with which this air penetrates the body under this pressure, very slight changes in it would prove disastrous to life. The variations in this pressure may arise in two ways, — first, through the changes in the air pressure at the same altitude (due to the passage of areas of high and low barometric pressure, in which the extreme cases would give a variation of not over 4 inches) from about 27.5 inches to 31.5 inches; second, through the change in altitude and increase in pressure by making a descent, or decrease in pressure by an ascent. In the former case of varying the pressure merely, a maximum change of 0.8 of an inch air pressure in a day would be equivalent to changing the altitude by about 690 feet during that time; or in the extreme case of an absolute change from 31.5 inches to 27.5 inches air pressure, it would be the same as coming from a depth of about 1,378 feet below the ground surface, and ascending to a height of about 2,133 feet above this surface. In other words, the changes in air pressure which we may experience at the earth's surface are as great as those which would be encountered in an ascent of about 3,500 feet in altitude.

There are, however, limits to the amount that the air pressure can be changed without harm. Just what these limits are, we do not know, and they would vary for different persons; but one individual (Humboldt) was exposed to a barometric pressure of about 48 inches in a diving bell, and to 14.75 inches on a mountain top; others have been exposed to a still lower air pressure.

It is probable that animals can stand a somewhat greater increase than decrease of pressure. In cases of such extreme variations of air pressure the changes must take place gradually. If they take place too rapidly, fainting, bursting of blood vessels, or even death, will result. If the change takes place slowly, then, as the limit of endurance is approached, certain physiological effects are made apparent. These

are the symptoms of the so-called mountain sickness, such as difficulty of breathing, headache, pressure on the eardrums, and a general apathy, which prevents exertion of any kind. These symptoms disappear, however, when lower altitudes are again reached. Permanent residence at very high altitudes does, however, impoverish the blood by decreasing the amount of oxygen inhaled in breathing.

Classified Distribution of the Air Pressures. — The whole lower air mass may be divided into regions of low and high air pressure, or areas of deficiency or excess, between which lie areas of more nearly normal pressure.

Regions of Low Barometric Pressure may be divided into four classes or orders: —

1. Depressions of the first order are those belonging to the air system of whole hemispheres. They include those permanent ones extending equatorward and poleward from about latitude 30°. They are of permanent character.

2. Depressions of the second order are of great extent, covering areas many hundreds of miles in diameter, and occur mainly in the middle and lower latitudes. They are not permanent as regards either location or duration.

3. Depressions of the third order are minor depressions occurring within depressions of the second order.

4. Depressions of the fourth order are those arising from the local excess of temperature. They are fixed in position, but are not permanent. Such occur during the daytime over limited isolated land areas (as, for instance, the Spanish Peninsula).

Regions of High Barometric Pressure may be divided into three classes or orders, as follows: —

1. Regions of high pressure of the first order are the permanent hemispherical ones to be found in the lower, middle, and north polar latitudes. They extend upwards to considerable altitudes.

2. Regions of high pressure of the second order are those resulting from the heaping-up of air between two depressions of the second order. They are therefore not permanent in either duration or location.

3. Regions of high pressure of the third order are caused by the local cooling of the air at the earth's surface. They are of local occurrence, and do not extend to great altitudes.

Barometric Gradient. — By *gradient* is usually meant the degree of steepness of a slope. In meteorology we speak of the steepness of the slope of isobaric surfaces as the barometric gradient. Instead of measuring the angle of the slope, it is customary to take the difference in barometric pressure along the same level. As a unit of distance in all directions, the length of a degree of the meridian is taken. The barometric gradient between two places is, then, the difference in the barometric pressures at the same level, divided by the number of meridional degrees between the two places. Thus, if at one place the barometric pressure is 30 inches, and at another place the pressure (reduced to the same level) is 30.50 inches, and the distance between the two places is five meridional degrees, then the barometric gradient between the two places would be $(30.5 - 30) \div 5$, or 0.1, of an inch.

Wherever a gradient exists, there is a gradient force acting, due to the efforts of gravity to render the isobaric surfaces level. The amount of this gradient force, in the case of isobaric surfaces, increases with increase of the magnitude of the gradient; that is, the steeper the inclination of the isobaric surfaces, the greater the gradient force. This gradient force is effective, or operates, in the direction of the slope of the isobaric surfaces, and air flows down a sloping isobaric surface in a manner somewhat similar to the flowing of water down an incline.

CHAPTER IV.

WINDS.

Air Currents. — When the air is at rest, relative to surrounding objects, we are scarcely aware of its existence; but when it is in motion, we notice that it exerts a pressure or force against these objects. Air in motion is called *wind.* When air is at rest, it is said to be calm. The air of the atmosphere would as a whole be continually calm, if it were not for the unequal distribution of temperature over the earth's surface and through the air itself. This unequal heating gives rise to air motion or *air currents*, the passage of which gives the phenomenon of wind.

The air in moving must go from some one place to another; and therefore wind has *direction*, as we see from the movement of light objects, such as leaves or clouds, which are carried along by the wind. There must also be a rate of motion, which we call *wind velocity;* and we judge of this by the rapidity with which the air transports such light objects. Heavy objects lag behind the true motion of the air.

Since air has density or mass, then when it is in motion it must exert a pressure or force, called *wind force* or *wind pressure*, against objects with which it comes in contact, as is realized by the way the wind bends trees, or blows against our bodies.

Air currents moving approximately parallel to the surface of the ocean are called *horizontal air currents.* Air

currents moving approximately perpendicular to the surface of the ocean are called *vertical air currents*. The possible movement of vertical currents is limited to a few miles between the earth's surface and the outer limit of the air, while horizontal air currents might make the entire circuit of the earth. The usual direction of the wind is some combination of these two directions. The vertical movement of the air is usually so slight, as compared with the horizontal movement, that ordinarily the latter only is observed for general winds.

The Winds of the Globe may be divided into the following classes: —

1. A permanent and continual interchange of air between the equatorial and polar regions, due to the differences in temperature between those regions: this is called the *general circulation of the atmosphere*. The annual shifting of the thermal equator (which follows the latitude of the sun) causes an annual displacement and inequality of this system of winds.

2. An interchange of air between the masses lying over bodies of water and of land, due to the unequal heating of the two surfaces. When the interchange takes place over great regions, as between continents and oceans, it has an annual period of occurrence: this is called the *monsoon wind*. When the interchange takes place between the coast lands and the coast waters only, it is diurnal in occurrence: this is called the *land and sea breeze*.

3. An interchange of air between mountains and valleys, due to the unequal heating and cooling of the two localities: this is called the *mountain and valley breeze*, and is diurnal in its occurrence.

4. An interchange of air between extensive regions of

higher and lower barometric air pressures: this gives rise to the *cyclonic* and *anticyclonic winds*, which are irregular in occurrence.

5. A local rush of air for the restoration of the normal condition (stable equilibrium) when the latter has been disturbed by local causes: to this class belong *spouts*, *tornadoes*, and *squalls*, which are of occasional occurrence.

6. A rush of air which is produced in front of avalanches and land slides: this is called *avalanche wind*, and is of occasional occurrence.

7. The outrush of air which takes place with volcanic eruptions: this may be called *volcanic wind*, and is of occasional occurrence.

8. To these must be added another class of winds, which may be called *eddy winds* or *whirlwinds*. These occur as follows: When a current of air is flowing steadily along, there break out at intervals from the sides whirls or eddies such as may be seen in flowing water; or they may be produced by the meeting of two air currents differing in direction. These are irregular in occurrence, and vary in magnitude from large whirls produced by the general air currents (No. 1) to the small dust whirls which we see in the streets.

Direction of the Wind. — The direction of the wind is named from the direction of approach. Thus, if a wind blows from the north, it is said to be a *north wind*; if it blows from the east, it is said to be an *east wind*; etc.

In accurate work in meteorology, 16 points of compass are used, as follows: —

North, north northeast, northeast, east northeast, east, east southeast, southeast, south southeast, south, south southwest, southwest, west southwest, west, west northwest, northwest, and north northwest. These are indicated by their initial letters, as shown in Fig. 29.

The Horizontal Direction of the Wind is usually observed by means of an instrument called the *arrow wind vane* (Fig. 30), the head of the arrow pointing in the direction of *approach* of the wind. The arrow is free to revolve horizontally around a vertical axis placed near the head. The best vanes have a divided tail with a divergence of about 22°.

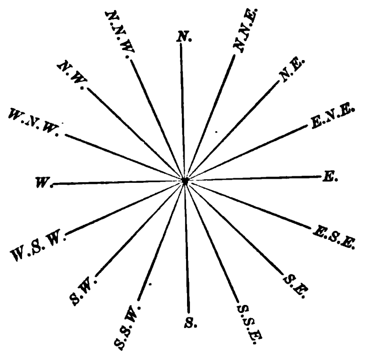

FIG. 29. — POINTS OF COMPASS FOR WIND DIRECTIONS.

The Velocity of the Wind is the rapidity with which the air moves past some fixed point, or covers the known distance between some two points. Wind velocities are usually given in miles per hour, or feet per second, in English measure.

The Anemometer. — An instrument used to measure the velocity of the wind is called an *anemometer*. It is a small wind wheel with an attachment by means of which the number of revolutions of the wheel is indicated on a little dial. When exposed to the wind, the rapidity of revolution of the wind wheel varies with the actual velocity of the wind. The commonest form is the Robinson cup anemometer shown in Fig. 31.

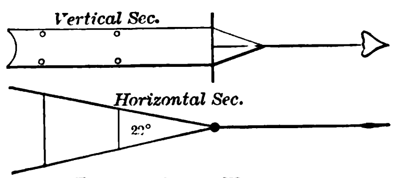

FIG. 30. — ARROW WIND VANE.

The four hemispherical cups revolve as a whole around a vertical axis, when exposed to the wind. It has been found that the distance traveled by any one of the cups, when multiplied by *about* 2.5, will give

the velocity of the wind. The distance passed over in one revolution of the cup is found by multiplying the distance from the center of the vertical axis to the center of the cup by (2) × (3.1416). The dial on the vertical axis shows the number of miles of wind.

The Force of the Wind is the pressure which it exerts on a flat surface held perpendicular to its direction of motion. It may be judged by an observer, or expressed in actual pounds of pressure per square foot of exposed surface.

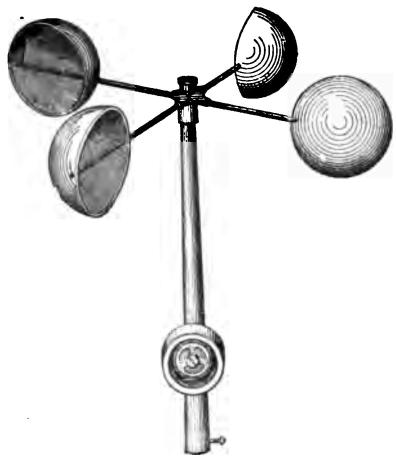

FIG. 31. — ROBINSON CUP ANEMOMETER.

Observation of Wind Direction.—In most cases when wind direction is observed two or three times daily, as in ordinary observations, the results are tabulated for each month and for the year by giving the number of times the wind blows from each of the eight principal points of the compass, and the number of calms during the time.

Thus in 1891 the number of times the wind blew from various directions at St. Paul during the year and the extreme months, as shown by two observations daily, was as follows: —

	N.	N. E.	E.	S. E.	S.	S. W.	W.	N. W.	Calm.
January	2	0	2	23	3	4	7	17	4
July	3	4	8	19	1	9	7	9	2
Year	24	39	57	224	43	90	84	113	56

The direction from which the winds blow with greatest frequency is called the *direction of prevailing winds.* It varies for different regions.

Tables have been published which give the relative frequency of the wind from the different points of compass, for the months and the year, for various regions of the earth; and charts have been drawn which give the direction of the prevailing winds, for each month and the year, for portions of the continents and traversed seas of the globe. (See wind directions given on the charts, Figs. 26, 27, showing the distribution of air pressure at sea level over the whole earth. The arrows fly with the wind.)

Where the wind blows at different times in different directions, all of these motions combined are equivalent to a motion in a single direction, called a *resultant direction.*

This may be explained graphically as follows: If we start at a point A, and draw a line in the proper direction equal to the amount of north wind, and from this end of the line draw another line in the proper direction equal to the amount of the northeast wind, and from the end of this line lay off in the proper direction another line equal to the amount of the east wind, and so on for all the wind directions, then a line drawn from A to the end of the last line drawn, will represent the direction, and in length the amount, of the resultant wind.

Observation of Wind Velocities. — Wind velocities are subject to diurnal and annual periodic variations and to marked irregular changes. The diurnal change of wind velocities has in general but a single maximum and minimum. The maximum usually occurs at about two to three hours after noon for the *average of the year* on the land near the surface of the ground, and about the same time before noon on the ocean; and the hour of minimum varies from two to eight hours after midnight on the land, but is nearer midnight (it may be before or after) on the ocean. The diurnal amplitude or difference between the maximum and minimum wind is small over the ocean, but over the

land it is large. In the United States it varies from 20 to 120 per cent of the average velocity of the wind.

The daily variations in the wind velocity at considerable altitudes above the ground are the reverse of those at the earth's surface as to time of occurrence of principal phases. From observations which have been made, it appears, that, at an altitude of perhaps 400 to 600 feet above the ground, the daily change in the wind velocity

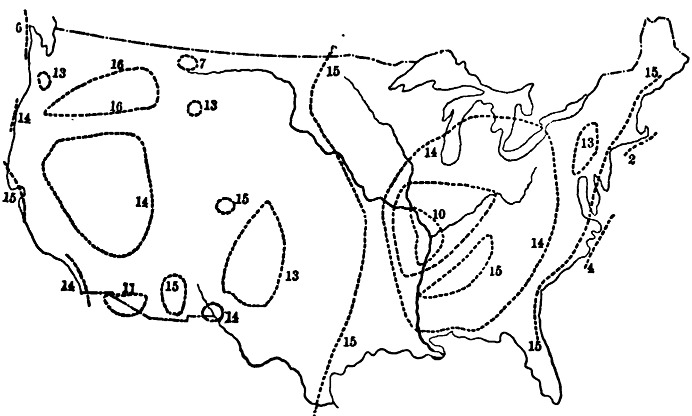

FIG. 32.—HOUR OF MAXIMUM WIND, IN THE UNITED STATES, FOR JANUARY.

practically disappears, while at greater altitudes occur phases the reverse of those at the surface of the ground, but with less amplitudes. In the warm season of the year the maximum and minimum phases are much more pronounced, and the time of maximum is later in the afternoon than in the colder season.

During the night there is little change in the wind velocity; but as soon as the sun's heat makes itself felt, there is a more or less rapid increase near the surface of the ground, and a slighter decrease at greater altitudes.

The cause of this has been explained as follows: **The wind velocities increase with the altitude above the ground to a certain extent.** When the ground is heated by the sun's rays, the warm air rises, and cooler air from above must descend to take its place. This warm air from the ground surface carries with it the lesser (horizontal) velocities at the surface, and mixing with the upper air retards its horizontal movement; while the descending upper air takes with it the greater horizontal velocities at the higher altitudes, and thus accelerates the horizontal movement of the lower air with which it mixes. Some of

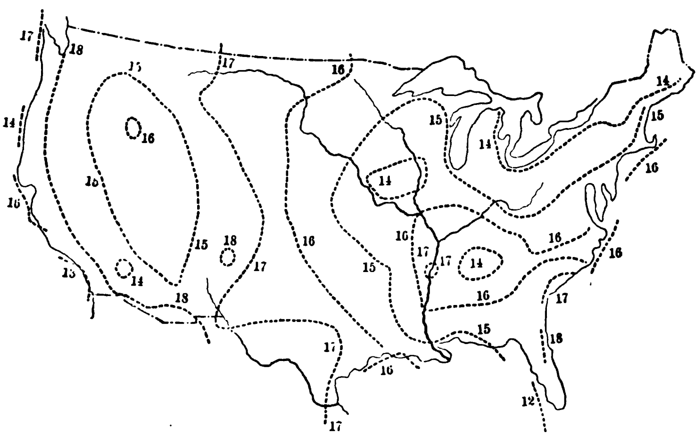

FIG. 33 — HOUR OF MAXIMUM WIND, IN THE UNITED STATES, FOR JULY.

the increased velocity of the lower air is directly due to the increased air movement caused by the warming of the air during the daytime, and would take place even if the velocities aloft were not greater than those below.

The accompanying charts (Figs. 32, 33) show the continental distribution of the hours of maximum wind for January and July.

Fig. 34 shows the daily march of the wind velocities, in miles per hour, for the extreme months of January and July at a few selected stations in the United States.

Great absolute diurnal variations in the wind velocity occur in some localities; as, for instance, in July for San Francisco and Corpus Christi (coast stations), and Whipple Barracks (an inland station).

On cloudy days the daily amplitude of wind velocity is relatively small, because the action of the sun is weak.

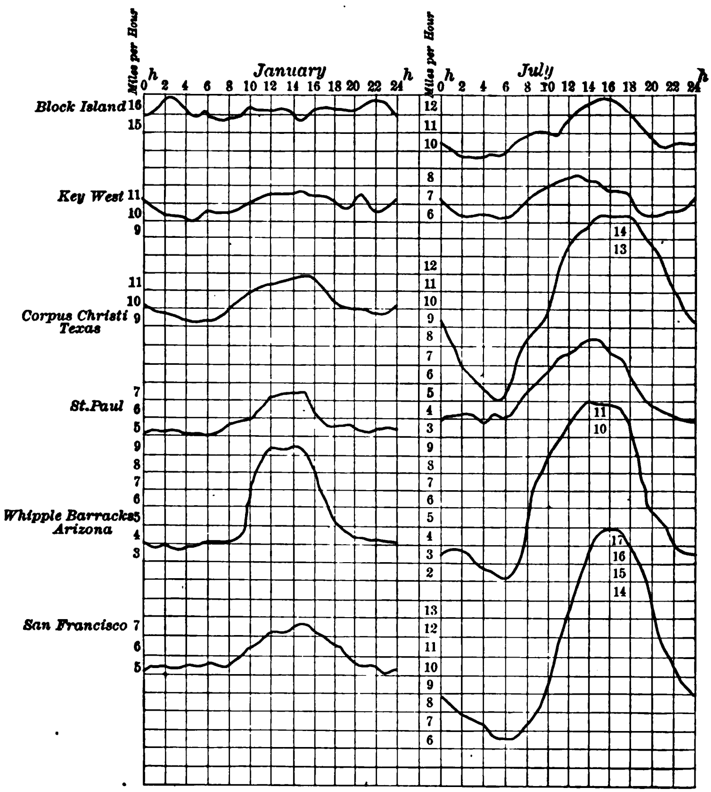

FIG. 34. — DIURNAL CHANGES IN WIND VELOCITIES (MILES PER HOUR).

The Annual March of the Wind Velocity has in general but a single maximum (usually during the cold season) and minimum (in the warm season).

The maximum occurs about March, and the minimum about August, in most of the United States; but in some portions of the

western section the primary maximum occurs about April, and there is a slight secondary minimum about December. In most of Europe the maximum wind occurs in a winter month.

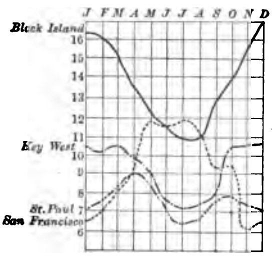

Fig. 35. — Annual Changes in Wind Velocities (miles per hour).

The Annual Amplitude of Oscillation of the average hourly wind velocities for the extreme months is variable for different regions, amounting to over 10 miles per hour in some cases; that is, the windiest month may have an average of over 10 miles per hour more wind than the calmest month.

In the United States the amplitude is about 4 or 5 miles per hour, but on the northwestern coast it is 11 miles per hour. On the central Great Plains there is relatively slight variation in the wind velocity for the different months of the year.

The following table and diagram (Fig. 35) show the average monthly wind velocities, and consequently the annual march of the wind, at selected places in the United States: —

Average Monthly Wind Velocities (*miles per hour*).

	J.	F.	M.	A.	M.	J.	J.	A.	S.	O.	N.	D.	Year
Block Island.	16.3	16.0	15.3	13.4	12.2	11.6	11.0	11.0	13.0	14.0	15.5	16.9	13.8
Key West.	10.4	10.2	10.6	9.8	8.9	7.6	7.3	7.4	8.0	10.7	10.7	10.7	9.3
St. Paul.	7.1	7.5	8.3	9.0	8.5	7.5	6.5	6.7	7.7	7.9	7.4	7.0	7.6
Dodge City, Kan. . .	10.3	10.5	11.6	13.3	12.9	12.1	11.1	10.4	11.2	10.7	9.5	9.7	11.1
Corpus Christi, Tex.	10.4	10.9	12.1	13.7	13.3	12.3	10.7	10.7	10.6	10.5	10.0	10.1	11.3
San Francisco	6.7	7.1	8.4	9.5	11.8	11.7	11.1	11.8	9.3	9.4	6.2	6.5	9.0
Tatoosh Island, Wash.	16.0	14.4	12.7	10.6	10.3	7.9	7.6	7.4	10.1	11.7	14.8	15.7	11.6

The Average Wind Velocity for the Whole Year has not been determined at enough places to permit the showing of lines of equal wind velocities for the whole earth. These have been drawn, however, for the United States

(p. 355) and for the Russian Empire, and roughly indicated for some portions of the oceans.

In the United States the wind velocities are greatest on the coast, and in general decrease towards the interior; but on the treeless Great Plains near the center of the continent there is an increase again to nearly the same velocities as are found along the low shores of the ocean.

In Russian Siberia there is a decrease towards the center of the continent; but there is no central region of increased wind, as found in the United States. Whether this is due entirely to the difference in the physical features of the two regions, or is partly due to the higher latitudes of Central Siberia, has not been determined.

The Increase of the Wind Velocity with Increase of Altitude is very rapid for the first hundred or two hundred feet; but above that it is slow, and very variable not only for the yearly averages, but also for the different months of the year.

It is quite probable that up to some unknown altitude the wind velocity increases, and at higher altitudes it decreases again. The increase in wind velocity with altitude is very much less over the ocean than on the land, for the first few hundred feet; but above this altitude there is probably not much difference between the wind velocities over a land and those over a water surface. The movements of clouds show that wind velocities of perhaps 200 miles per hour sometimes occur at high altitudes.

Obstacles to Air Motions. — Air movement does not progress unhindered, for the moving air layers near the ground rub against it, and those up above rub against other air layers which have a different direction of motion, or no motion at all. This rubbing is called *friction*, and it retards the air currents.

The friction of one current of air on another is exceed-

ingly slight; but the friction of the air against the earth is very great.

The main hindrance to the air currents acquiring enormous velocities is the continual mixing or mingling of air layers having different directions of motion, and especially the breaking-up of the air into vortices, within which the air layers which had the initial velocities become so increased in numbers, and separated, and so broken up into strata which twist spirally around one another, that the friction, and especially the interference of such innumerable surfaces having inequalities of motion, becomes very great, and the motion is thus equalized. Mountain ranges may also somewhat retard the velocities of the lower air currents.

The Decrease of Wind Velocities through Friction with the Ground is well shown by comparing the wind velocities on the land, where the friction and direct obstacles are greatest, with those on the ocean, where they are least.

It is found by observation that the wind velocities over the land at the height of low buildings (40 or 50 feet above ground) are about 25 %, at the height of the tallest buildings (100 to 150 feet above ground) about 50 %, and on the well-exposed seashore about 75 %, of the normal velocities existing on the open ocean (40 feet above the water) for the same geographical locality.

The Variation of Lower Wind Velocity with the Latitude is not easy to determine in amount, on account of the different exposure of anemometers; but it is thought that in the northern hemisphere, at least, there is an increase in wind with increase of latitude up to latitude 50°–60°, and then a decrease farther poleward.

The discussion of a few observations on the open Atlantic Ocean showed an increase, in the average wind for the whole year, of from about 8 miles per hour in latitude 22° or 23° north, to 17 miles per hour

at about latitude 50° north, from which latitude northward to 58° north there was again a slight decrease; but how this decrease continues onward towards the pole has not been determined.

The general region of maximum wind in middle and higher latitudes is shown for the Atlantic Ocean in Figs. 36 and 37, which give for January and July the wind relations of the Atlantic Ocean as regards both the direction, and roughly the force, of the wind. The heavier arrows denote the stronger wind, and the double arrows the strongest wind; short arrows denote variable, and long arrows steady wind; circles show region of prevailing calms. These show in a general way the increase in wind velocities from the tropics up to the higher middle latitudes.

The Periodic Daily Change in the Wind's Direction is relatively slightly marked, but still it is sufficient to have been recognized. It stands in close connection with the daily variation in velocity, since the vertical air currents occurring about midday and in the early afternoon communicate not only part of their velocity, but also their direction, to the lower air currents. The air currents above have a tendency to move or deviate to the right of the lower air currents, and this motion causes in the northern hemisphere, over a level land surface, a tendency for the wind in the morning to turn in the direction of the hands of a watch or movement of the sun, and towards evening to turn in the opposite direction; but in the higher atmosphere (as on high mountain peaks) this is reversed. In the southern hemisphere, on the level land surface, the morning turning is in the direction opposite to the motion of the hands of a watch, and towards evening with it; and in the upper atmosphere this is reversed.

At the equator the vertical air currents do not cause this diurnal turning of the wind's direction.

On the ocean the tendency for the wind's direction to turn as described is very slight on account of the weak

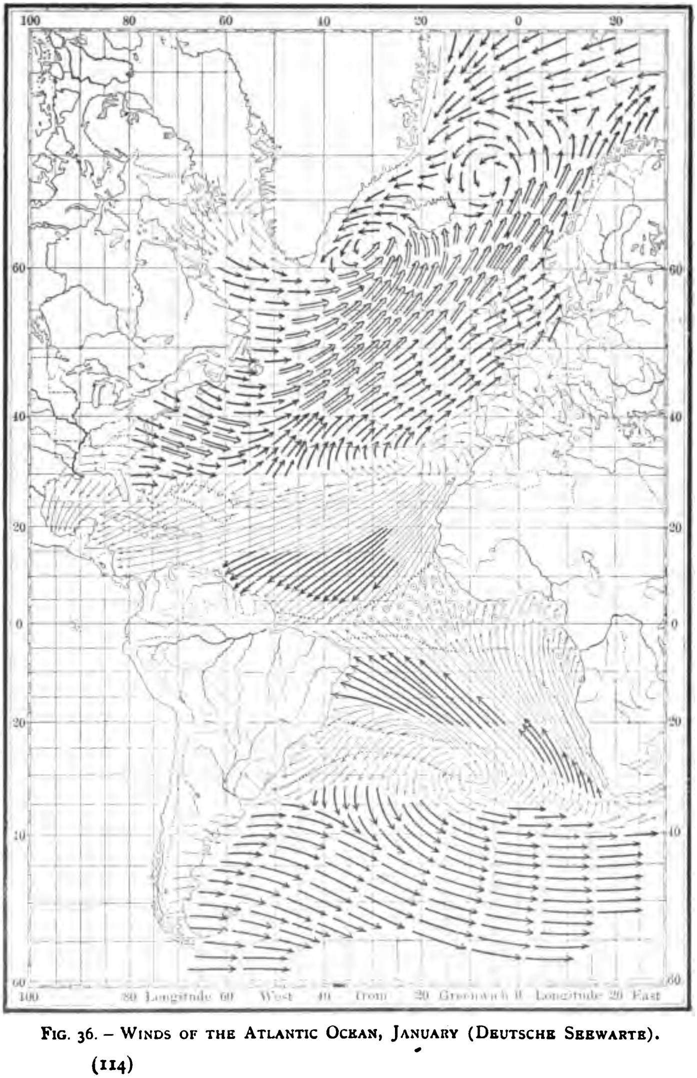

Fig. 36. – Winds of the Atlantic Ocean, January (Deutsche Seewarte).

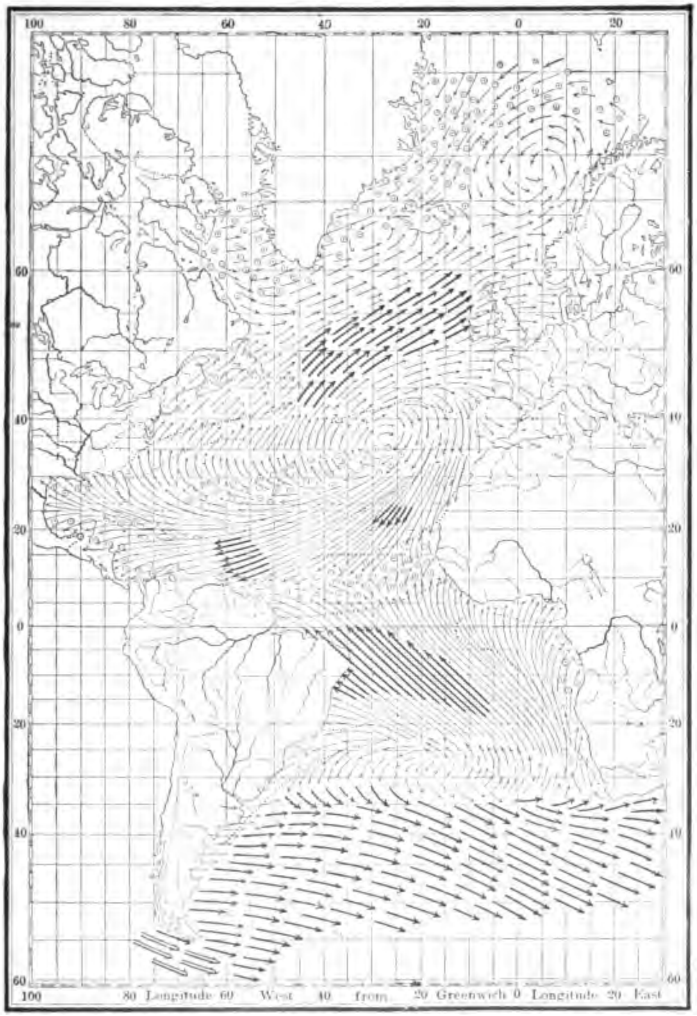

FIG. 37. — WINDS OF THE ATLANTIC OCEAN, JULY (DEUTSCHE SEEWARTE).

vertical currents which arise over the relatively cool water in the daytime.

The local diurnal changes in the wind's direction (known as sea and land breezes, and mountain and valley winds) are described in treating of the atmospheric motions (p. 262).

The Periodic Annual Changes in the Wind's Direction are due to the seasonal shifting of the temperature conditions which have already been mentioned.

First, there is the change in the general circulation of the atmosphere, due to the shifting of the region of greatest heat, first to the north, and then to the south, of the equator.

Secondly, there is an inflow of the lower air towards the center of the continents in summer, and an outflow from the continents in winter. These are called the *monsoon winds;* and they cause reversals of the direction of the general flow of air in the regions affected.

Thirdly, there is a seasonal variation in the general courses or tracks of barometric minima which move across the earth's surface, and this changes the general direction of the wind in the region lying in their paths.

In the middle latitudes, where these annual changes are most marked, the direction of the wind is a most important factor in determining the climate of the region.

The resultant wind directions at selected stations are, —

	Jan.	Feb.	March	April	May	June
St. Paul . .	S. 73° W.	S. 70° W.	N. 62° W.	N. 10° W.	S. 75° E.	S. 4° E.
Key West .	N. 59° E.	N. 66° E.	N. 76° E.	N. 87° E.	N. 82° E.	S. 65° E.

	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
St. Paul . . .	S. 31° W.	S. 4° E.	S. 2° W.	S. 37° W.	S. 82° W.	S. 79° W.	S. 33° W.
Key West . .	S. 71° E.	S. 76° E.	N. 88° E.	N. 60° E.	N. 54° E.	N. 53° E.	N. 78° E.

The resultant direction of the horizontal and vertical components of the wind has never been determined with any great accuracy, but it must on the average amount to only a few degrees from the horizontal.

Formation of Air Waves. — It is seen that when the wind blows over a water surface there results the well-known phenomenon of water waves. Similarly, if one current of air moves over another at rest or of a different velocity or direction, and the line of demarcation between the two is sharply drawn, there will be a favorable condition for the formation of air waves. Such conditions do occur very frequently in our atmosphere, and since the effects of friction between two air currents moving one over the other are so very slight, we can have very great differences in velocity existing between the two within a short distance of the boundary between them. Air waves have a magnitude 2,500 times that of water waves produced by the same wind conditions. The summits or crests of air waves are made visible by some of the detached clouds high up in the sky; and their troughs are made evident by the gusts of wind felt in their passage near the surface of the ground.

The further discussion of the winds will be taken up in other chapters on the subject of the movements of the atmosphere.

CHAPTER V.

MOISTURE: VAPOR, CLOUD.

Moisture of the Air. — There is a continually repeated interchange of water going on between the air and the surface of the earth, and in considering the circuit through which the water passes in these changes it is most convenient to separate it into three sections; viz., —

1. The moisture as it exists in the atmosphere as invisible vapor and as cloud, or the *hygrometry of the air.*

2. The rate at which it condenses and descends to the earth as *precipitation,* and the amount of this precipitation; or the rain, hail, and snow fall.

3. The rate at which the moisture is taken up again into the air from the earth's surface, or the *evaporation.*

Moisture exists, either as a gas or vapor, in an invisible form; or as a liquid or solid, in visible form. The change from one form to another depends mainly on the conditions of temperature. For a given temperature, only a certain maximum amount of moisture can exist as a vapor. When this temperature is reached by cooling, saturation occurs, and any further cooling condenses the moisture, and renders it visible.

Evaporation is that process by which a liquid is converted into its vapor; as, for instance, the conversion of water into water vapor. Some solids also give off vapor, ice among others. The molecules of both the liquid and

its vapor are in constant motion. The average velocities in the vapor are greater than in the liquid; but the largest velocities in the liquid may be greater than the smallest in the vapor. If any of the molecules at the surface of the liquid have these abnormally large velocities, and if they are moving *from* the liquid, they will escape from those forces which retain in the liquid the molecules of lesser velocity, and will fly off from the liquid as vapor. The number of molecules which pass from the liquid to vapor depends mainly on the temperature, but partly also on the amount of vapor already existing above the liquid.

In the process of evaporation there is a disappearance of a quantity of heat, and a depression of temperature occurs. This heat which disappears is called the *latent heat of vaporization*. Its amount varies according to the temperature at which evaporation takes place. For a further account of this, the reader must consult a text-book of physics.

Condensation. — If the molecules of vapor have their velocities diminished, by cooling, down to the velocities of the liquid, then condensation occurs, and the vapor is converted into a liquid; or the molecules of vapor striking the liquid may become entangled among the molecules of the liquid, and thus become part of the liquid. The number of molecules which pass from the vapor to the liquid depends on the density of the vapor as well as on the temperature.

Saturation. — If the temperature of the vapor and liquid is the same, evaporation will be in excess of condensation until the density of the vapor has become so great that as many molecules are condensed as are evaporated, then the vapor has attained its maximum density (for that temperature), and is said to be saturated. Evaporation and condensation are still going on, however, after saturation; but they just balance each other.

Conditions of the Atmosphere with Regard to Moisture.— Four conditions of the atmosphere with respect to moisture are recognized, which may be termed *stages*, since they are brought about by a progressive succession of changes to which the moist air is subjected.

There is first the *dry stage*, in which the air contains vapor; but it is not saturated, and consequently no condensation or precipitation takes place. The amount of unsaturated or superheated vapor in the dry stage for the free air (during its changes) may be assumed to remain constant; that is, no vapor is added or abstracted. In this dry stage, if the air mixture has been cooled, then by the addition of heat, or by adiabatic compression, the original condition can be resumed, as no water has been lost.

The *rain stage* is reached when the unsaturated moist air of the dry stage is cooled by the abstraction of heat, or subjected to adiabatic expansion until the condition of saturation is reached and passed, and water drops form.

The water drops fall away from the air as rain, unless, by means of a strong upward wind, they are carried along with the air in which the condensation occurred. The absolute amount of moisture in the air is decreased by the amount which falls away as rain, and the remaining vapor is in a saturated condition.

In the rain stage the changes which occur are seldom reversible, that is, capable of returning the air to its original condition, as has been mentioned for the dry stage, because usually some of the moisture becomes lost as raindrops. In the rain stage there are two extreme conditions which may exist: either all of the water drops which have been formed may fall away as rain, or the water drops may be carried along with the air from which they were condensed, by means of the support offered by sufficiently powerful vertical air currents. In the former case the temperature will fall more rapidly than in the latter as the change goes on.

The *hail stage* is reached when th... and cooling moist air reaches the freez... The air mixture in this stage is made ... rated vapor, any water drops which are p... water-drops which exist as ice or hail. Wh... ture is just at the freezing point, water-drop... lets may exist side by side.

At the beginning of the hail stage we ha... the rain stage, in which the air mixture contained water in the form of raindrops and saturated vapor. The amount of hail will depend on the amount of these water drops when the hail stage was entered upon, and also on the amount of change in the volume of the mixture by expansion. It must be carefully noted that the hail stage cannot be entered upon until there is an appreciable amount of water present as raindrops.

The *snow stage* is reached by the gradual cooling of the ascending air to a temperature below freezing of water, but without the formation of raindrops; or when the precipitation of rain and hail is so slight that the latter stage, at least, is practically skipped over. In the snow stage the air mixture consists of dry air, saturated vapor, and snowflakes, and the mass is still less than for the hail stage.

Since the amount of vapor which may be present depends on the temperature, then, the lower the temperature, the less the possible snowfall; which accounts for the fact of comparatively light snowfalls in extremely cold weather.

By a reversion of the process by which these stages are reached — that is, by a direct warming, or by adiabatic compression of the air mixture — it is possible to carry it back into the dry stage again; but the amount of vapor

less than that it started with, by the amount of moisture which has fallen away in the form of rain, hail, or snow.

Atmospheric Moisture as Vapor.

The Amount of Moisture in the air existing as a vapor may vary from nothing up to saturation. These two extreme conditions serve as terminal points for a scale of measurement of the degree of moisture in the air. The temperature is the most important factor in determining the amount of moisture, and in general the possible maximum amount of moisture diminishes with decrease of temperature; or, the colder the air, the less the possible amount of moisture.

There are three terms constantly used in speaking of atmospheric moisture: *absolute humidity*, *relative humidity*, and the *dew-point*.

Atmospheric Humidity. — The humidity of the air is the amount of moisture which it contains; that is, the amount which is mixed with the dry air to form the atmospheric air.

The Absolute Humidity is the weight of this moisture. If we consider it in its watery condition, it is then expressed in grains per cubic foot of air. If we consider it in its invisible gaseous condition, then we measure the pressure or tension of the vapor, and it is then expressed in fractions of an inch of barometric pressure.

The Relative Humidity is the relation of the amount of moisture present to the amount necessary for saturation under the existing condition, and it is expressed in percentage of the latter. When it is said that the relative humidity is 50 %, this means that half as much moisture is present as would be necessary for the saturation of the

vapor under the existing conditions of temperature and barometric pressure.

The Dew-point. — Since the amount of vapor which can exist in the air depends chiefly on the temperature of the air, then, if we make any moist air colder, we shall increase its relative humidity; and if it is cooled far enough, the relative humidity will become 100 %, and saturation will occur.

The dew-point is that temperature of the air at which its invisible moisture begins to condense into visible water drops.

The following table shows the weight, in grains Troy, of saturated vapor per cubic foot at the annexed dew-point temperatures: —

Dew-point, F°	0°	40°	60°	80°	100°
Grains Troy	1.3	2.9	5.8	11.0	19.8

If a little ice is put into a tin cup of water, and the water is stirred so that all of the water may cool at the same time, and if the bulb of a thermometer is put into the water and the temperature is noted when the moisture first begins to be deposited on the outside of the cup (like the "sweating" of a pitcher of cold water on a warm day), then the thermometer reading gives the dew-point of the air for its existing conditions. It is a little difficult to detect the exact instant at which the condensation of the moisture on the outside of the cup begins, and so it is usual to note as accurately as possible the temperature when this condensation begins, and then to warm the water slightly and note its temperature when the condensed moisture disappears; and the average of the two temperatures will give quite closely the desired dew-point.

Methods of Measuring Atmospheric Moisture. — There are several methods of determining the amount of moisture in the atmosphere, but those in chief use by meteorologists are: (1) by determining the temperature of evaporation (by using the psychrometer); (2) by observing the proportional saturation of certain animal and vegetable substances (by using the hair hygrometer); (3) by observing the various aspects of the sky.

Determination of Humidity by the Temperature of Evaporation. — The *psychrometer* is used in this method. This instrument consists of two similar mercurial thermometers placed side by side; the one being in its unaltered condition, and the other having a wet thin muslin covering for the bulb which contains the mercury. The muslin cloth is usually made wet by connection with a wicking dipped into water. The evaporation of the water from the cloth cools the thermometer bulb to the temperature of evaporation; and the drier the air, the lower will the temperature of the wet bulb sink below that of the dry bulb. A psychrometer is shown in Fig. 38. More accurate results than for still air are obtained by making an air current pass over the thermometer bulb, which can be done most easily by whirling the thermometers.

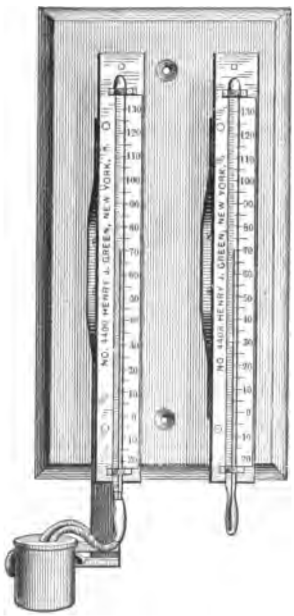

FIG. 38. — PSYCHROMETER: WET AND DRY BULB THERMOMETERS.

The Hair Hygrometer (Fig. 39) is sometimes used for determining the relative humidity of the atmospheric air. This instrument is based on the fact that hair expands in length with increase of moisture, and contracts with decrease of moisture. The amount of this expansion or contraction is measured on a scale which shows the percentage of total expansion for saturation.

Actual Observation of Atmospheric Humidity. — The atmospheric relative humidity most directly affects our

sensibilities, and it is generally expressed in per cent of saturation. For many questions of physical investigation concerning the atmosphere, however, it is necessary to know the absolute amount of water in the atmospheric air, expressed either in vapor pressure or in weight of water.

The Daily March of the Relative Humidity is very pronounced. In general it decreases with the diurnal increase of temperature. There is an early morning maximum and an afternoon minimum. During the daytime, when the temperature increase is rapid, just before and about noon, the relative humidity falls very rapidly, but during the night it does not vary by a great amount. Where there is a large daily amplitude for the oscillation of temperature, there will also be found a great amplitude for the relative humidity; and where the amplitude of the temperature is small, that of the relative humidity will likewise be small. The diurnal amplitude of the relative humidity is small near the ocean coasts, but increases towards the interior of the continent.

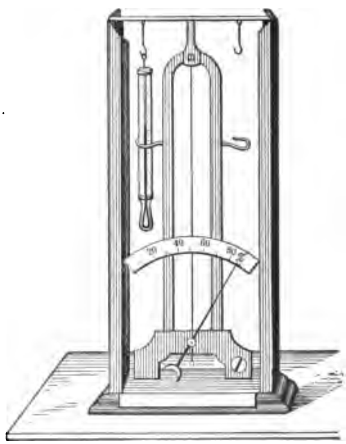

FIG. 39.—HAIR HYGROMETER.

On the northwest coast of Europe the amplitude is 7% in winter (December), and 17% in summer (August); while at Nukuss, in central Asia, the amplitude is 26% in winter (December), and over 50% in summer (August).

The relative humidity has the greatest amplitude on clear days, and the least on cloudy days. The amplitude is several times as great in

clear weather as in rainy weather. There is a tendency for the relative humidity to increase on the average for the 24 hours of cloudy days, and to decrease during clear days.

The Annual March of the Relative Humidity is somewhat irregular. The time of maximum is in midwinter, and of minimum in early summer. The annual amplitude is least at the coast stations, and greatest in the interior of continents. The amplitude for the extreme months on the coast of the ocean amounts to perhaps 5 %, and in the interior of the continent to over 30 %.

The average annual relative humidity is greatest for the marine climate, and decreases towards the interior of the continents. Up to a certain altitude (that of the clouds), the annual amount increases with the elevation, but beyond that point it decreases. But it varies greatly with altitude, since the temperature and amount of vapor, upon which relative humidity depends, vary so much with altitude.

Local influences, such as mountains, the direction of the winds with regard to the neighboring drier or moister regions, and other causes, affect the relative humidity of a place.

The annual march of the relative humidity at a few stations is shown in percentage by the following average monthly relative humidities: —

	J.	F.	M.	A.	M.	J.	J.	A.	S.	O.	N.	D.	Year
Key West, Fla. . .	80	77	70	69	71	71	70	72	75	76	78	79	74
St. Paul, Minn. . .	72	71	69	60	60	68	70	72	71	69	72	74	69
Salt Lake City, Utah	60	57	58	45	40	32	30	31	31	42	52	60	44

The Geographical Distribution of the Relative Humidity is not easily shown on charts, on account of its variability with altitude. On the open ocean, however, it is quite

regular, and on the average varies between 82% at the equator and 92% in high latitudes. On the continents, owing to the high temperatures, there is usually a small *relative* humidity in summer, and a large one in winter; but for the *absolute* humidity we have just the reverse of this.

The Daily March of the Absolute Humidity varies greatly for different places. There are, however, two pronounced types, — the *maritime* and the *continental.*

The *maritime* type has a maximum in the warmest part of the day, with an early morning minimum. The *continental* type has two maxima, one just before noon, and the other late in the afternoon, with a secondary minimum between; the primary or lowest minimum occurring in the early morning.

On the mountains the amplitude is less than on the plains. The secondary minimum in the continental type is due to the ascending air currents which occur towards the hottest part of the day; the surface air being carried upward more rapidly than the moisture is supplied to it by the evaporation which takes place from the earth's surface. Over the water surface these ascending currents are not so strong as over the land (especially on dry plains), and therefore the air is not so likely to rise upward faster than the lower layers are supplied with moisture by evaporation.

The absolute humidity has the greatest amplitude on clear days, and the least on cloudy days, when precipitation also occurs. There is a tendency for the absolute humidity to decrease during the 24 hours of clear days, and to increase during cloudy days.

The Annual March of the Absolute Humidity is very similar in character to that of the temperature. The minimum is in the winter months, and the maximum in the midsummer months. The amplitude for the year is least on the coasts, and greatest in the interior of con-

tinents; and it is less in the tropics than in the temperate zones.

The Average Amount of Vapor decreases with the altitude above sea level, and it decreases more rapidly than the decrease of air pressure (and density) with altitude.

The relative decrease of the amount of water vapor with the altitude, and also the decrease of the air density, are shown by the following table, in which the air density and water vapor at the earth's surface are each placed at 1.00:—

ALTITUDE.	WATER VAPOR.	AIR DENSITY.
Feet.		
0	1.00	1.00
13,000+	.24	.61
30,000−	.04	.32

It is seen how much more rapidly the water vapor decreases than the air density. Half of the water vapor lies below the altitude of about 6,500 feet, while half of the air lies below the altitude of about 18,000 feet. The low altitude of the greater portion of the moisture in the atmosphere accounts in a great measure for the powerful influence which even low mountains exert in the distribution of this moisture over the earth's surface.

The Geographical Distribution of the Water Vapor over the different regions of the earth follows quite closely that of the temperature; in general the greater the temperature, the greater the amount of water vapor. With increase of latitude the amount of vapor decreases. In January the region of greatest amount of vapor is the region extending from the equator to about latitude 20° south.

In this region the vapor pressure is over 0.8 of an inch, but in equatorial Africa it reaches 1 inch, and on the northern coast of Australia 0.95 of an inch. In western Europe it varies from 0.2 to 0.4 of

an inch, in eastern Europe from 0.1 to 0.2 of an inch, and in the cold regions of northeastern Asia (and likewise in north-central North America) it decreases to 0.05 of an inch or lower.

In July the region of highest vapor pressure lies to the north of the equator, where in India it reaches 1 inch.

In the extreme northern part of the northern hemisphere the vapor pressure is reduced to 0.2 of an inch or less; while in the southern hemisphere the pressure of 0.2 of an inch is reached even at latitude 40° south.

It is thus seen that there is a movement of the maximum region backwards and forwards across the equator from one hemisphere to the other, following the sun.

Atmospheric Moisture as Cloud and Fog.

Fog and Cloud Formation. — When air is cooled just below the dew-point, then fog or cloud occurs, and, as may be seen by observing the sky, in a great variety of forms. While in a general way the process of cloud formation is understood, yet the exact manner of the building of clouds of the various shapes and appearances which occur has not been entirely studied out. Clouds are at present divided into classes, according to the forms which they present to the eye, and not according to their processes of formation.

Composition of Clouds. — Since the clouds contain more particles of dust than the surrounding clear air, it is concluded that the air within clouds has been drawn up from close to the earth's surface, where these dust particles abound. The dust particles within clouds usually number only a few thousand per cubic centimeter; but the water particles may number as high as 50,000 in the same space.

The density of a cloud depends directly on the number of water particles present.

When cloud forms, precipitation begins at once; but the water particles are very small, and evaporation takes place when they fall into drier air below. The distance which they fall depends on their size and the dryness of the air beneath. The denser the cloud, the larger are the raindrops, and the faster they fall.

Nomenclature of Cloud Forms. — There are four distinct classes of clouds, according to their forms, — *cirrus, stratus, cumulus*, and *nimbus*.

Cirrus clouds are those which are seen in striated forms grouped high up in the sky.

Stratus clouds are those which present a stratified or bank-like form. They may be high or low lying clouds.

Cumulus clouds are more or less isolated, have rounded tops, and are found at but moderate altitudes above the ground.

Nimbus clouds are those from which rain or snow descends.

Combinations of these forms add several varieties with recognized names. The principal ones are as follows: cirrus, cirro-stratus, alto-stratus (strato-cirrus), stratus, cirro-cumulus, alto-cumulus (cumulo-cirrus), strato-cumulus, cumulus, cumulo-nimbus, nimbus. These various forms are represented in the accompanying illustration (Fig. 40).

1. *Cirrus clouds* (No. 1) are feathery in form and delicately fibered, usually of a white color, and well outlined against the sky background They lie arranged in a variety of fantastic forms. Nearly parallel groups of these clouds are sometimes seen stretching across the heavens in converging, meridian-like bands. Such bands as these are also sometimes formed of cirro-stratus and cirro-cumulus clouds. The cirrus clouds have perhaps an average altitude of between 5 and 6 miles.

2. *Cirro-stratus clouds* (No. 2), consisting of fine, white, veil-like clouds, and the *alto-stratus* clouds (No. 5), thick, veil-like clouds of grayish or bluish color, look in some respects quite alike in form, but their altitudes are very different; the altitude of the cirro-stratus being about 5½ miles on the average, and that of the alto-stratus being only about half as great, perhaps about 3 miles. The cirro-stratus clouds usually precede bad weather, and they gradually give place to alto-stratus clouds. The cirro-stratus are the clouds which give rise to the phenomenon of rings around the sun and moon. The cirro-stratus formation is sometimes very widely diffused over the heavens, and then becomes a cirrus vapor; but sometimes a distinct but intricate-fibered structure is visible, which may be called *cirrus-felt*.

3. *Cirro-cumulus clouds* (No. 3) are very small white, and *alto-cumulus* (No. 4) large whitish-gray, balls or fleecy clumps grouped in herds. Sometimes these fleecy clumps are arranged in rows, extending in one or two directions. The cirro-cumulus lie much the higher, being at an altitude of 3¾ to 4½ miles, while the alto-cumulus are only about 2½ miles up.

4. *Strato-cumulus* (No. 6) and *nimbus* (No. 7) *clouds* belong to the lower air layers, and form clumps or layers at an altitude of from ¾ of a mile to 1¼ miles above the ground. Strato-cumulus clouds occur in dry weather, and appear very frequently in winter, when they more or less cover the heavens, but with patches of blue sky between the clouds. The nimbus is the cloud of continued rain or snow. It has ragged edges, and above are always to be seen the alto-stratus clouds; so that between the nimbus clouds a gray cover or background is visible, and not the blue sky such as is seen above the strato-cumulus.

5. *Cumulus clouds* (No. 8) are the thick, dense clouds with rounded, festoon-like tops and horizontal bases. The top usually comes to a point or peak higher than the rest of the cloud. These clouds are built up by ascending currents within; and when these air currents cease, the clouds disappear gradually. Such clouds are characteristic of the summer sky, especially over the land. Sometimes over tropical islands the vertical air currents which arise cause more or less permanent cumulus clouds to overhang the islands.

Cumulo-nimbus clouds (No. 9) are the thunder and shower clouds which roll up in such an imposing manner, and present a majestic appearance of mountain-like character. The tops are of a light, fluffy appearance, while the bases are of the dense nimbus character, from whose

center showers of local rains and hail descend. The upper edge is festooned like the cumulus, and towering cumulus peaks are formed. The edges are sometimes bordered by small, cirrus-like clouds, or these last are detached from the main cloud.

6. *Stratus clouds* (No. 10) consist of an elevated fog. Fog which lies at the ground is designated simply *fog;* but when it is at an altitude of, say, a thousand or more feet above the ground, it is called *stratus cloud.* Such is dry-weather fog.

The broken, tattered clouds which appear at low altitudes in wet weather are called *fracto-nimbus.* The tattered clouds which occur with the true cumulus clouds are called *fracto-cumulus.*

Processes of Cloud Formation. — The following processes are those by which fog and cloud are produced: —

1. Direct cooling of the moist air through contact with colder bodies or through loss of heat by radiation.

2. Adiabatic cooling of ascending moist air.

3. Mixture of moist air of different temperatures and humidities.

These are mentioned in the order of their effectiveness for the production of cloud.

Dissipation of Clouds. — Conversely, fog or cloud may be dissipated in the following manner: —

1. Direct warming of the clouded air through radiation or contact with warm bodies.

2. Adiabatic heating of descending clouded air.

3. Mixture of the clouded air with other air masses of proper temperature and humidity.

Ground Fog. — The condensation produced by direct contact with a cold body and by loss of heat through outward radiation is that of ground fog, which extends upwards to a moderate height above the ground. The outward radiation from the ground on clear nights cools the ground very rapidly; and, as soon as the dew-point is passed, condensation begins in the lowest air layers, and fog forms. The

upper surface of this fog radiates heat, and cools the air layers just above it, so that condensation ensues in these layers. Thus the process of fog building proceeds, and the fog layer becomes thicker and thicker. When the solar radiation begins to make itself felt, the reverse takes place; and the upper layers are dissipated first by the warming of their upper surface; then the next layer is dissipated; and so on until the ground is reached, when it too becomes warm, condensation ceases, and the fog entirely disappears.

The reason that no excessive precipitation occurs in this process is that the formation of the fog cloud above the ground prevents the further excessive cooling of the latter by continued radiation. The fog growth on the upper limit of elevated clouds may also occur in the manner just described; but, in order to produce condensation by direct radiation in the upper air layers, there must exist a cloudiness formed by some other process, or by means of a framework of such impurities as smoke particles.

The Formation of Clouds through Adiabatic Expansion, and their dissipation through compression, occur where there exist ascending and descending air currents. The huge summer clouds with rounded tops and horizontal bases, the so-called thunderclouds, and the usual rain clouds, are formed by this process, when ascending air currents are present, and the air thus expands and cools; and their dissipation occurs for descending movements, whereby the air is compressed again and made warmer.

The Formation of Cloud by Mixture of Air of Different Temperatures and Humidities is a much more complex matter than that just described; and for cloud formation it is of great importance, but is comparatively unimportant for causing the precipitation of moisture to the ground.

Condensation takes place more rapidly when a current

of cool moist air penetrates a large mass of warm moist air, than when a current of warm moist air enters a mass of cool air.

Condensation does not always proceed gradually, and it may not occur until the final stages of cooling by mixing are reached, when it will take place all at once. Likewise in the reverse process the dissipation of the cloud may be retarded until the condition is such that the whole of the cloud will disappear quite suddenly. In the case of the breath leaving the nostrils and penetrating the cold air, the condensation does not take place until actual mixture is effected; and when this mixing process distributes the breath further through a larger mass of the drier colder air, the fog is dissipated. It has been shown, that, in the mixing of saturated cool air with larger quantities of saturated warm air, the warming of the former takes place at first rapidly, and then more gradually; but when the cold air is greatest in quantity, then the cooling of the warm air becomes more rapid as the process continues. The maximum amount of condensation will occur when the amount of cool air is in excess.

By this process of mixing, the following kinds of clouds and fog are produced: —

1. The fog which arises over relatively moist warm surfaces, where cooler air is also present. An example of this is the surface ocean fog which so frequently arises during the cold season.

2. The clouds which are formed at the common boundary of two air currents (the air in which may differ as regards both temperature and moisture) moving with different velocities, by which means a regular succession of clouds is formed as a consequence of a wave motion which may exist; and so great is the amplitude of these waves, that the adiabatic condensation arising during the upward surges must make itself visible.

3. The layers of stratus clouds which are formed at the juncture surface of two such air currents having different velocities, and which have at first a disconnected form, but which afterwards become joined.

4. The banner-like clouds which form (and dissolve) on some mountain peaks, and in mountain passes, when the contour is such that warm or cold masses of air are penetrated by currents of air having other temperatures.

5. The loose and tattered clouds which are so noticeable in strong winds, and especially in thunderstorms, and which are constantly undergoing changes of form.

It is very probable that fogs and clouds of various forms are also produced by combinations of any two or all three of the causes of condensation; but we do not as yet know their method of forming well enough to specifically point them out. It may also be remarked, that, without a knowledge of the mode of formation of clouds, it is a very easy matter to misinterpret their movements as we observe them.

Amount of Cloud. — The degree of cloudiness is estimated on a scale of from 0 when there are no clouds visible, to 10 for the whole sky overcast with clouds. It is therefore merely estimated how many tenths of the sky are covered with cloud. Almost all of such observations are made by the unaided eye. These estimations are reduced to percentage of the total visible sky by multiplying them by 10.

While there has been some slight attempt at estimating the amount of the various kinds of clouds, yet in nearly all cases the degree of cloudiness is estimated irrespective of the kind, density, or height of the clouds. Thus, if the degree of cloudiness is recorded as 7 (on a scale of 0 to 10), and cirrus and stratus clouds are present, no attempt is made to state what proportion of this 7 refers to the cirrus and what to the stratus clouds.

The Daily March of Degree of Cloudiness is somewhat difficult to determine, since the amplitudes are small, and vary only about 10 % of the whole surface of the sky. In the morning hours fog and stratus clouds are the most frequent; but towards midday, and in the afternoon, the cumulus are in excess. In general there is a principal maximum of cloud about noon or a little after, and a principal minimum in the night. This varies, however, not only with change of locality, but also at different seasons of the year.

The Annual March of Degree of Cloudiness shows in general a maximum amount of cloud in the late fall or early winter, and a minimum in the spring, but sometimes retarded until summer for relatively low lands; but in the high lands (the Alps, for instance) the maximum occurs in the summer, and the minimum in the winter.

The maximum cloudiness occurs mostly in early winter, because then the average temperature is decreasing rapidly, which causes an increase in the relative humidity. The minimum occurs mostly in the spring, because then the average temperatures are increasing rapidly, and this causes the relative humidity to decrease. The average monthly cloudiness depends, however, so much on the direction of the winds with regard to the supply of moisture, the relative humidity, and other conditions, that a great variety of phases are to be met with.

The amount of cloud in Bombay varies from 18 % in December to 91 % in July. In Cairo, Egypt, it varies from 33 % in December to 6 % in June. In Spitzbergen it varies from 51 % in December to 87 % in September. On the northern coast of Norway it varies from 68 % in January to 60 % in June.

The following table shows the average monthly amount of cloud, and consequently the annual march of cloudiness, at a few places in the United States (in percentage of total cloudiness): —

	J.	F.	M.	A.	M.	J.	J.	A.	S.	O.	N.	D.	Year
Key West, Fla. . . .	42	35	29	31	43	48	49	49	52	46	38	40	42
St. Paul, Minn. . . .	49	48	51	51	52	49	42	45	48	51	58	51	50
Salt Lake City, Utah .	54	52	53	52	46	31	30	30	26	40	47	56	43
Prescott, Ariz. . . .	28	26	29	26	17	13	32	33	17	16	17	25	23

The Variation of Average Cloudiness with Latitude is shown by the following table, in which the cloudiness is given in percentages of the whole sky visible at one time: —

AVERAGE CLOUDINESS.

LATITUDE.	NORTHERN HEMISPHERE.		SOUTHERN HEMISPHERE.	
	JANUARY.	JULY.	JANUARY.	JULY.
70°	55	59	—	—
65°	56	61	—	—
60°	62	62	70	70
55°	59	61	70	67
50°	57	57	70	62
45°	52	50	60	54
40°	50	44	53	56
35°	46	41	48	52
30°	44	42	47	47
25°	37	45	48	46
20°	37	50	51	43
15°	40	53	53	48
10°	45	59	58	51
5°	50	59	56	57
Equator 0°	50	58	—	—

The following conclusions are drawn from charts made of the cloud distribution. The degree of cloudiness is arranged in zones parallel to the equator. The zones of minimum cloudiness are reached at 20°–25° north latitude, and about 25°–30° south latitude, and the amounts are least in the northern hemisphere. The region of maximum cloudiness is reached at about latitude 60° in the northern hemisphere, and then there is a decrease toward the north pole; but in the southern hemisphere there is an increase up to 60° south latitude, which is as far as observations extend. These zones present a well-defined tendency to follow the sun in its variations in latitude for the year, lying at a higher latitude in summer than in winter.

Fogs. — Fog at the earth's surface shields the ground from the solar rays in proportion to its density; and at its upper limit it must acquire

a temperature something like that which would occur at the earth's unclouded surface. The thickness of a fog is shown by the length of time necessary for it to be dissipated by the solar rays. Above the ground fog the air is clear and dry, and radiation takes place unhindered. Thus it happens that bright clear weather is experienced after the dissipation of a fog, which at first seems to be a forerunner of rain.

Ocean Fogs. — When warm moist air is carried by the atmospheric circulation into a cooler region, the moisture is condensed, and fog forms.

Such cases occur with marked frequency over the North Atlantic Ocean in the region to the south and east of Newfoundland, and consequently directly in the track of the transatlantic ocean steamers. In storms, when the wind blows from the east or southeast, the warm moist air from the Gulf Stream is blown over the colder waters of the Arctic current or over the ice floes which this current brings down from the north, and fog condensation takes place. The greatest frequency of fog occurs in those latitudes at about the 55th meridian (west of Greenwich). The number of days with fog for each month in that region is as follows:—

J.	F.	M.	A.	M.	J.	J.	A.	S.	O.	N.	D.	Year.
5	11	12	15	18	17	23	22	15	13	10	4	165

The sea air contains more moisture in the summer time than in winter, which, together with the great differences in temperature due to the large quantity of field ice and icebergs off Newfoundland, causes the relatively great frequency of fogs during the warm season.

Maximum Amount of Water in the Air. — The largest possible amount of moisture which can exist at any one time in the air (without the supporting power of upward air currents) between the earth's surface and certain altitudes is best shown by the number of inches' depth of rain which it would make if all of the moisture were condensed, and fell as rain. In the most favorable case there would not be

two inches of water between the earth's surface and an altitude of three miles above it.

Height of Column of Air above the Ground.	Depth of Water for the Following Dew-points at the Earth's Surface.			
	80° *F.*	70° *F.*	60° *F.*	50° *F.*
Feet.	Inches.	Inches.	Inches.	Inches.
6,000	1.3	1.0	0.7	0.5
12,000	2.1	1.5	1.1	0.8
18,000	2.5	1.8	1.3	0.9
24,000	2.7	2.0	1.4	1.0
30,000	2.8	2.1	1.5	1.1

The above table shows the amount of water which may exist in the air below certain altitudes for various temperatures.

center showers of local rains and hail descend. The upper edge is festooned like the cumulus, and towering cumulus peaks are formed. The edges are sometimes bordered by small, cirrus-like clouds, or these last are detached from the main cloud.

6. *Stratus clouds* (No. 10) consist of an elevated fog. Fog which lies at the ground is designated simply *fog;* but when it is at an altitude of, say, a thousand or more feet above the ground, it is called *stratus cloud*. Such is dry-weather fog.

The broken, tattered clouds which appear at low altitudes in wet weather are called *fracto-nimbus*. The tattered clouds which occur with the true cumulus clouds are called *fracto-cumulus*.

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upper surface of this fog radiates heat, and cools the air layers just above it, so that condensation ensues in these layers. Thus the process of fog building proceeds, and the fog layer becomes thicker and thicker. When the solar radiation begins to make itself felt, the reverse takes place; and the upper layers are dissipated first by the warming of their upper surface; then the next layer is dissipated; and so on until the ground is reached, when it too becomes warm, condensation ceases, and the fog entirely disappears.

The reason that no excessive precipitation occurs in this process is that the formation of the fog cloud above the ground prevents the further excessive cooling of the latter by continued radiation. The fog growth on the upper limit of elevated clouds may also occur in the manner just described; but, in order to produce condensation by direct radiation in the upper air layers, there must exist a cloudiness formed by some other process, or by means of a framework of such impurities as smoke particles.

The Formation of Clouds through Adiabatic Expansion, and their dissipation through compression, occur where there exist ascending and descending air currents. The huge summer clouds with rounded tops and horizontal bases, the so-called thunderclouds, and the usual rain clouds, are formed by this process, when ascending air currents are present, and the air thus expands and cools; and their dissipation occurs for descending movements, whereby the air is compressed again and made warmer.

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6. *Stratus clouds* (No. 10) consist of an elevated fog. Fog which lies at the ground is designated simply *fog;* but when it is at an altitude of, say, a thousand or more feet above the ground, it is called *stratus cloud.* Such is dry-weather fog.

The broken, tattered clouds which appear at low altitudes in wet weather are called *fracto-nimbus.* The tattered clouds which occur with the true cumulus clouds are called *fracto-cumulus.*

Processes of Cloud Formation. — The following processes are those by which fog and cloud are produced: —

1. Direct cooling of the moist air through contact with colder bodies or through loss of heat by radiation.

2. Adiabatic cooling of ascending moist air.

3. Mixture of moist air of different temperatures and humidities.

These are mentioned in the order of their effectiveness for the production of cloud.

Dissipation of Clouds. — Conversely, fog or cloud may be dissipated in the following manner: —

1. Direct warming of the clouded air through radiation or contact with warm bodies.

2. Adiabatic heating of descending clouded air.

3. Mixture of the clouded air with other air masses of proper temperature and humidity.

Ground Fog. — The condensation produced by direct contact with a cold body and by loss of heat through outward radiation is that of ground fog, which extends upwards to a moderate height above the ground. The outward radiation from the ground on clear nights cools the ground very rapidly; and, as soon as the dew-point is passed, condensation begins in the lowest air layers, and fog forms. The

upper surface of this fog radiates heat, and cools the air layers just above it, so that condensation ensues in these layers. Thus the process of fog building proceeds, and the fog layer becomes thicker and thicker. When the solar radiation begins to make itself felt, the reverse takes place; and the upper layers are dissipated first by the warming of their upper surface; then the next layer is dissipated; and so on until the ground is reached, when it too becomes warm, condensation ceases, and the fog entirely disappears.

The reason that no excessive precipitation occurs in this process is that the formation of the fog cloud above the ground prevents the further excessive cooling of the latter by continued radiation. The fog growth on the upper limit of elevated clouds may also occur in the manner just described; but, in order to produce condensation by direct radiation in the upper air layers, there must exist a cloudiness formed by some other process, or by means of a framework of such impurities as smoke particles.

The Formation of Clouds through Adiabatic Expansion, and their dissipation through compression, occur where there exist ascending and descending air currents. The huge summer clouds with rounded tops and horizontal bases, the so-called thunderclouds, and the usual rain clouds, are formed by this process, when ascending air currents are present, and the air thus expands and cools; and their dissipation occurs for descending movements, whereby the air is compressed again and made warmer.

The Formation of Cloud by Mixture of Air of Different Temperatures and Humidities is a much more complex matter than that just described; and for cloud formation it is of great importance, but is comparatively unimportant for causing the precipitation of moisture to the ground.

Condensation takes place more rapidly when a current

The effect of increase of rainfall with altitude is very curious in some dry regions where at low altitudes only a small amount of precipitation takes place; and at higher latitudes on the mountain sides there will be such an increase of rainfall as to form belts of land where enough rain is received to cause a considerable or even luxuriant growth of vegetation in an otherwise arid region. Above this watered region there may be such a decrease in the rainfall as to make it impossible for plants to grow.

The Geographical Distribution of Annual Rainfall. — It has been estimated that about 6% of the land surface has an average annual rainfall of over 75 inches; 16% has from 50 to 75 inches; about 25% has from 25 to 50 inches; over 30% has from 10 to 25 inches; and over 20% has less than 10 inches. The results of rainfall observations on the ocean are unsatisfactory, owing to the lack of permanent places of observation. The distribution of rainfall (precipitation) over the land surface is well shown on the accompanying chart (Fig. 42).

The darkest shading shows where the rainfall is over 75 inches. The greatest rainfall far exceeds this amount, however, for a number of localities. Among the Chassia Hills in India the average rainfall is over 470 inches, and in some other localities in the tropics the amount reaches 190 inches. On the tropical islands, and especially those containing elevated land, the rainfall is usually excessive; also on the portions on the west coast of Europe where there are high ranges of hills, on the west coast of North and South America beyond latitude 40°, and on the eastern coasts of Newfoundland and Japan, the rainfall is tropical in amount.

Copious rainfall, that is, between 50 and 75 inches, is shown by the next lighter shading. This is the rainfall for the greater part of Central Africa and South America, regions of smaller extent in southeastern Asia and in southeastern and northwestern North America, northwestern Europe, portions of central Europe, and the eastern coasts of Australia and central Asia, and the higher eastern coast of North America.

The region of moderate rainfall, that between 25 and 50 inches, shown by still lighter shading, covers most of the eastern half of North

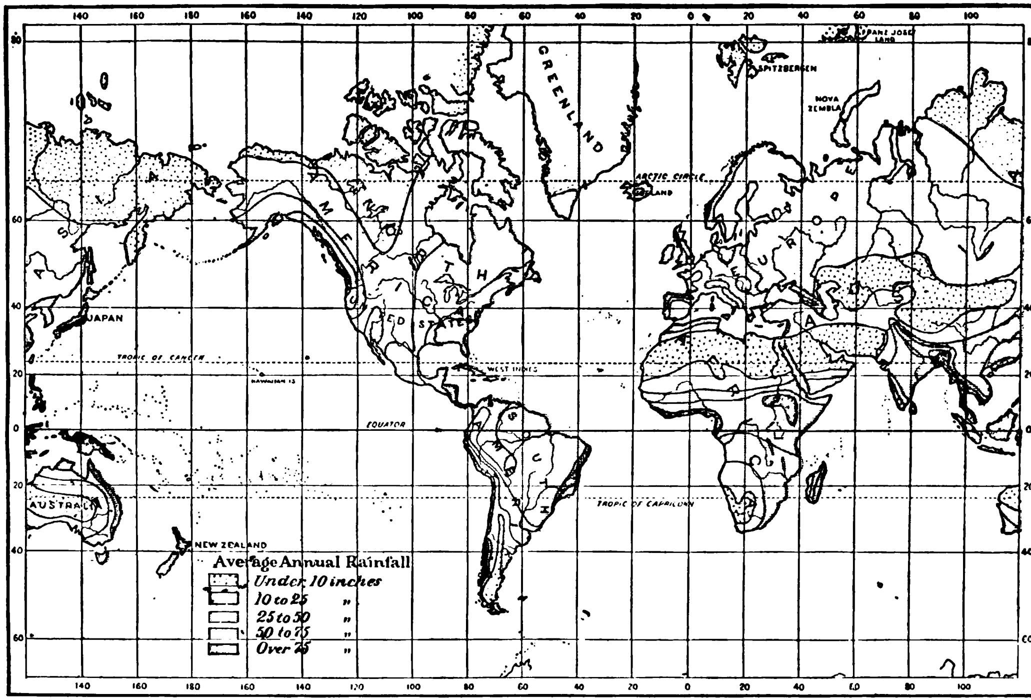


FIG. 42.—DISTRIBUTION OF ANNUAL RAINFALL (PRECIPITATION) OVER THE LAND SURFACE (AFTER LOOMIS).

America, central Europe, southeastern Asia, and south-central Africa, South America, and Australia. Northern Africa forms a transition zone to regions of less moisture.

The region of light rainfall is that having between 10 and 25 inches of rain. This covers an enormous territory in the northern hemisphere, the most of the western half and much of northern North America, northern Asia, and Russia; and it forms broad transition zones to the arid regions to be next mentioned.

The arid regions, those in which the rainfall is below 10 inches, are found in north-central and southwestern North America; south-central and west-central South America; southwestern and almost the whole of northern Africa; southwestern, central, and northeastern Asia; and central and western Australia.

The general seasonal rainfall distribution over the globe is shown on the accompanying chart (Fig. 43). Within each of these broadly characterized regions there are minor subdivisions having special features, but presenting great complexity of detail. These have therefore been omitted on the chart. The major subdivisions are as follows: —

1. (Light Pink) Region of Tropical Rainfall: Principal dry season in winter and spring; maximum rainfall in summer.

2. (Dark Blue) Region of Subtropical Rainfall: Maximum in winter; summer rainless.

1.2. (Dark Purple) Transition Region between 1 and 2: Rain winter and summer.

3. (Medium Blue) Region of Maximum Rainfall in winter; somewhat less rain in summer.

4. (Dark Pink) Region of Minimum Rainfall in late summer, with considerable rainfall at other seasons.

5. (Light Purple) Region of Moderate Rain during all the months. (Snow in winter.)

6. (Light Blue) Wet Region: Copious Rains during the whole year, but most rain in winter.

7. (White) Dry Region: All months of the year deficient in rainfall.

Characteristics of Rainfall in the Distinctive Rain Regions of the Earth. — The rainfall areas have been divided into the subequatorial or tropical, subtropical, and temperate regions. The tropical region lies on each side of the equator, between the tropics. The subtropical region extends mostly

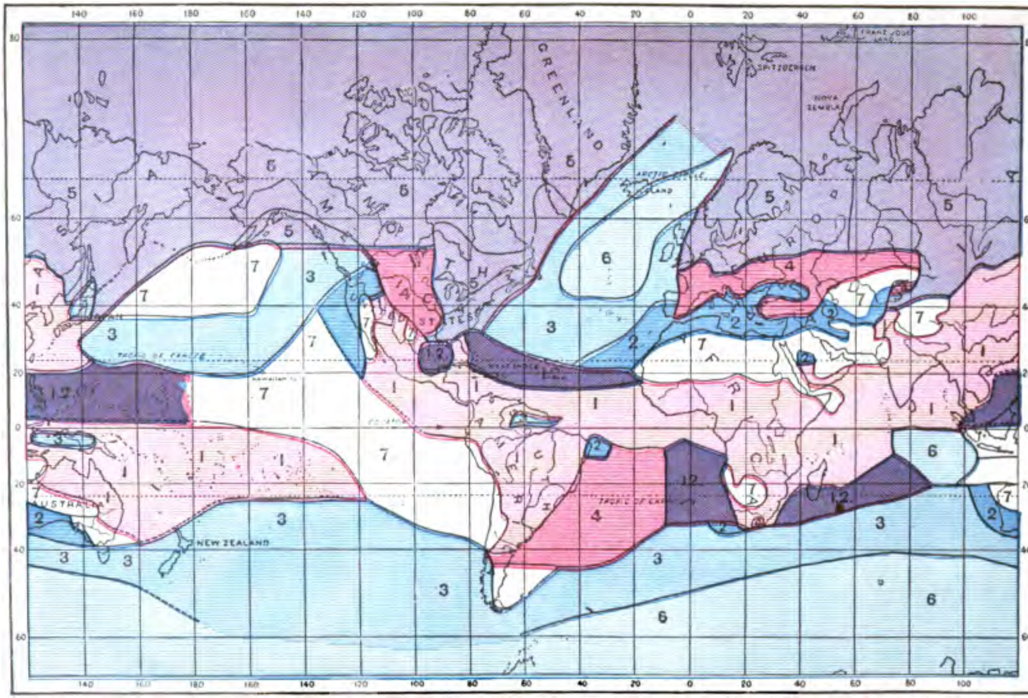


FIG. 43.—RAINFALL SEASONS OF THE GLOBE (AFTER VAN BEBBER).

from about the tropics poleward, and reaches as far as the 40th parallel in many cases. The temperate region lies to the poleward of the tropical region.

The Tropical Rainfall occurs mostly in summer, and is partly the accompaniment of local storms, and due to the vertical air currents which arise especially at the time of greatest midday heat in summer, and is partly due to the steady winds from the east bearing the moisture-laden sea air on to the mountains of continents and islands. Also, as in India, the monsoon winds bring a rainy season when they blow from the ocean landwards; but it requires a range of mountains to produce summer rains in the path of these steady air currents.

There is a shifting of the tropical or subequatorial rain belt with the shifting of the sun from one hemisphere to the other, because with this there is also a shifting of the equatorial wind system. In the region of calms (the doldrums) occur mainly the rains due to the local upward currents; and the shifting of this region follows the annual course of the sun in the heavens. In the region over which the doldrums pass twice in the year there are two rainy seasons, with two dry seasons between.

The excessive rainfalls of the eastern coasts and mountain slopes near the equator are due to the steady winds carrying the moisture from the oceans to the land; but where the excessive rainfalls occur on the western coasts, they are due to the action of vertical air currents in local storms, because, when the winds have passed over the continents from the east, they have been deprived of much of their moisture. The amount of water usually existing in the air as vapor is very great in the equatorial region, owing to the high temperatures prevailing there.

Long-continued series of rainfall observations have been made at but few places in the tropical rain belt, outside of India, but the following table shows the average monthly and annual rainfall at some selected stations. These show the rainfall conditions on both the eastern and western coasts of the continents.

TROPICAL OR SUBEQUATORIAL RAINFALL (*in inches*).

AFRICA.	LATITUDE	J.	F.	M.	A.	M.	J.	J.	A.	S.	O.	N.	D.	Y.
Loando (west coast) .	9° S.	2.4	1.1	1.3	3.3	0.3	0.0	0.0	0.0	0.1	0.2	2.5	1.2	12.4
Gaboon " . .	0°	7.5	9.0	15.0	8.7	8.9	0.9	0.5	1.2	8.1	19.4	19.1	7.5	105.8
St. Louis " . .	16° N.	0.2	0.5	0.0	0.0	0.2	0.4	2.6	8.0	3.7	0.5	0.0	0.0	16.1
Port Louis (east coast)	2° S.	5.7	11.8	5.2	3.1	2.1	1.5	0.9	1.5	0.4	0.7	1.7	3.7	38.3
TROPICAL AMERICA.														
Rio Janeiro (east coast)	23° S.	5.4	4.7	5.9	3.4	4.8	1.5	1.3	2.8	3.3	3.9	5.7	5.2	47.9
Georgetown "	7° N.	6.9	5.8	7.3	7.3	14.1	13.9	11.0	7.4	2.6	2.5	5.6	10.8	95.1
Havana "	23° N.	3.3	1.6	1.5	3.2	4.1	5.7	4.9	4.8	6.0	6.7	2.2	2.2	46.3
Quito (west coast) . .	0°	3.2	6.4	4.4	6.6	5.1	2.5	1.4	1.8	1.8	4.4	6.2	2.8	46.6
ASIA.														
Singapore	1° N.	8.4	6.1	6.8	7.0	6.3	7.2	6.2	9.1	7.1	8.8	10.7	10.8	94.5
Hongkong	22° N.	0.4	1.5	2.6	3.7	9.5	17.2	14.2	12.5	13.9	5.7	2.9	0.6	84.7
Calcutta	22° N.	0.4	1.0	1.3	2.3	5.6	11.8	13.0	13.9	10.0	5.4	0.6	0.3	65.6
Cherrapungi (Assam)		0.6	2.6	9.0	29.6	50.0	110.0	120.5	78.9	57.1	13.6	1.8	0.3	474.0

It is seen that in most cases the excessive rainfall occurs during but a few months of the year.

The Subtropical Rainfall is characterized by a winter rainy season and summer dryness. The total rainfall is relatively small, and is from three to ten times as much during the colder half year as during the warmer six months. In some of the regions (as, for instance, in Egypt and Palestine) there is scarcely any rainfall during the summer months. This region stretches up to nearly the 40th degree of latitude over the ocean, but is very variable on the continents, owing principally to the modifying circumstances of mountain ranges and winds. In the neighborhood of the Mediterranean Sea there is a decrease in rainfall from north to south, and from the west towards the east.

The following table shows characteristics for this region, where the lands bordering on the Mediterranean Sea give the best examples of the type: —

RAINFALL IN SUBTROPICAL REGION (*in inches*).

	J.	F.	M.	A.	M.	J.	J.	A.	S.	O.	N.	D.	Y.
Laguna, Teneriffe . .	9.6	5.7	6.1	2.2	1.3	0.4	0.0	0.0	0.4	2.2	4.8	10.9	43.6
Madeira	6.4	2.9	2.6	1.5	1.2	0.6	0.0	0.3	1.2	2.6	5.5	4.4	29.2
Sahara	1.0	1.1	1.7	1.4	1.3	0.6	0.2	0.5	1.2	1.4	0.8	1.0	12.2
France (south coast)	2.5	2.0	1.7	1.7	2.2	1.0	0.5	1.0	3.0	4.0	3.4	1.7	24.8
Italy (south coast) .	3.1	2.5	2.8	2.8	2.2	1.3	0.6	1.6	2.5	3.8	4.1	4.1	31.5
Beirut	5.4	7.2	4.3	3.3	0.7	0.4	0.0	0.0	0.7	1.8	4.7	7.6	36.2
Alexandria	2.1	1.7	1.0	0.1	0.0	0.0	0.0	0.0	0.1	0.2	1.7	1.9	8.7

The subtropical rains occur in that region in which the annual shifting of the atmospheric circulation with the course of the sun brings into play in the summer the dry trade winds blowing from the east; and in the winter the general winds from the west, which are rain-bearing, because in that great western current occur rain-producing storms (the cyclones of lower middle latitudes, to be described later). When the trade winds from the east are present, the season is dry; when the winds from the west are present, the season is wet. The stormy winds from the west make the west coast in this region very wet during the rainy season, because they bring the moist air from the ocean to the land. On the east coast the winds from the east may still be the most rainy, because they bring the moist air to the land, and the west winds have been deprived of much of their moisture in crossing the continent.

Rainfall of the Temperate Region. — In the temperate region, which is to the poleward of the subtropical,

there is usually rain in all the months of the year, but the amount is not usually distributed evenly over all seasons.

The rainfall causes are very complex in this region. It is the region of the greatest number of storms which occur in the strong permanent air current from the west. These storms are most frequent in the cold season, which increases the amount of rainfall at that time of year. In winter time, too, winds blowing the moisture-laden air from the oceans landward, carry it from a warmer to a colder region; and, the temperature of the air being lowered, moisture is condensed, and falls as rain. In the summer time the interiors of the continents become greatly heated, and violent local storms arise, which cause an increase of summer rains.

In western Europe the least rainfall occurs in midsummer, and the greatest in the fall or winter; in central Europe the minimum occurs in the early spring or winter, and the maximum in summer; in Russia and Siberia the maximum occurs in summer, and the minimum in winter.

In North America, on the Pacific coast, there is a winter maximum of rainfall, and an almost rainless summer; for the Mississippi River valley there is a summer maximum, and a minimum in the winter (but sometimes in the spring or fall); for the eastern coast of the United States there is a late summer maximum and early summer minimum for the middle regions, a winter maximum and early summer minimum at the north, and at the south a late winter maximum and late summer minimum.

In South America, on the west coast, there is a winter maximum and summer minimum; in the interior this is reversed; and on the east coast there is a winter minimum.

On the South African coast there is a winter maximum and a summer minimum, but in the interior there is an autumn maximum and winter minimum.

In Australia, on the eastern coast, there is a summer or early autumn maximum and winter minimum, and in the interior a very irregular rainfall. In south and west Australia there is a winter maximum and early autumn or summer minimum.

In the northern Arctic regions there is a late spring or early summer minimum and a fall maximum, except north of Asia, where there is a winter minimum and summer maximum, as for Siberia.

The Intensity of Rainfall is obtained by dividing the total rainfall by the number of days on which rain falls. The intensity is calculated for each month and for the year. The distribution of the intensity of rainfall for the months follows more closely the amount of rainfall than it does the number of rainy days. In normal cases the intensity is greatest in the warm seasons, but this does not always so occur.

Duration of Rainfall. — The average number of hours per day during which rain falls is quite variable for different regions, not only for the year, but for the different months.

The greatest average duration of the rainfall occurs in the cold season of the year in normal cases in the middle latitudes. On the coast of Norway the average number of hours is 11; in southeastern England, 5; in the northeastern quarter of the United States, 5; in the southeastern quarter, 4; in the dry region in southwestern United States, 2.5; and in the dry Rocky Mountain region, about 3.

The Variability of the Rainfall is obtained by dividing the average deviation of the individual cases from the average rainfall, by the average amount of rainfall. The variability

of rainfall (expressed in percentage of the average amount) in general increases with decrease of the absolute amount of rainfall. In regions of moderate rainfall (30 inches per year) 50 years of observations may leave a possible error of from 5% to 10% in the average annual rainfall.

The variability in the number of days of rainfall in the year, and for different months, is found to be the least for regions having the fewest number of days with rainfall, and greatest for those having the greatest number of days with rainfall.

The occurrence, during the same months, of the maximum or the minimum phases of rain amount, rain frequency, and rain intensity, is rarely found in the continental regions of the temperate zone.

The Probability of Rainfall is obtained by dividing the number of rainy days by the number of all the days during a chosen period, as a month or a year. This is quite an important element to be considered concerning the rainfall. The variation is so great from region to region, or even within narrow limits, that of course the phases of maximum and minimum frequency cannot be pointed out unless special mention is made of a great number of regions. It does not necessarily always happen that the greatest frequency of rainfall occurs with the greatest amount of rainfall.

Long-Period Fluctuations of Rainfall. — There is probably a periodic oscillation in the amount of rainfall extending over a period of 35 years. During about 17 years the rainfall increases slightly in amount, and then decreases again through the same length of time.

We have record of rainfall observations going back over 200 years at some places. The records for this century are more accurate and more numerous than those for the last century, and show for the land

the following periods of excess and deficiency of rainfall above and below the average amount: 1815, excess of rainfall; 1831–35, deficiency; 1846–50, excess; 1861–65, deficiency; 1876–80, excess. What the corresponding conditions were over the oceans, we do not know.

The Amount of Oscillation of the Rainfall during the probable thirty-five year period is of as great interest as the times of oscillation. The average oscillation is best expressed in terms of the average amount of rainfall. In Europe the oscillation is about 15 %; in Asia, 30 %; in North America, 25 %; in Central and South America, 25 %; the average for all the regions being about 25 %, or one fourth of the average total rainfall.

It has been found that intensity of the oscillations increases towards the interior of the continents. Thus in central Siberia 2.3 times as much rain falls in the period of excess of rain as in the period of deficiency, while in England it is but 1.2 times as much. Observations along the coast of the Atlantic Ocean have shown that there is a tendency towards a minimum of rainfall at the period of maximum rainfall in the interior of the continent; and this suggests that on this ocean, at least, there may exist a compensatory rainfall oscillation the reverse of that for the land.

There is, during these long-period oscillations of rainfall, a shifting back and forth of the rainfall areas, or, as the matter may best be expressed, a shifting of the lines of equal rainfall. A rough computation of this shifting of the regions of equal rainfall oceanwards and landwards shows that the region of 24 inches rainfall lies in Euro-Asia 1,000 miles, and in North America 700 miles, farther inland in the period of excess of rainfall than it does in the period of deficiency. The oscillations in the dry regions of Siberia are enormous. The rainfall which occurs in West Siberia in the relatively wet period is only to be met with at a distance of 2,000 or 3,000 miles from that region during the relatively dry period. The oscillations are least on the coast, and increase towards the interior of continents.

The Permanent Increase or Decrease of Rainfall in special localities, due to the increase or decrease in the forests, has

been the subject of much investigation, and it has not as yet been satisfactorily settled. Experiments for the artificial production of rain by means of explosions high in the air are not likely to be successful, as no appreciable amount of rain could be produced in this manner, unless ascending air currents could be caused and maintained for some time.

Hail. — When raindrops become frozen in their passage through the air, they fall as hail. Sometimes the raindrops may be frozen in their downward passage; but it is believed that they are usually frozen by being first carried upward by vertical air currents into regions where the temperature is below freezing, and that they then descend to the earth before they have time to melt. Sometimes hailstones have a soft snowy center; and frequently they have several successive coats or shells of ice, probably due to their being subjected a number of times to alternate temperatures above and below freezing.

Snow. — When condensation takes place at a temperature below freezing, minute ice crystals instead of water globules are formed, and the union of these crystals gives us snowflakes. Snow crystals assume a great variety of beautiful hexagonal forms, some of which are shown in the accompanying figure (Fig. 44). While the condition of freezing is necessary for the formation of snowflakes in the air above, yet they frequently, by falling rapidly without melting, reach the ground when the temperature of the lower air is much above the freezing point of water.

Latitudinal Limit of Snowfall near Sea Level. — The temperature of the *lower* air increases from a temperature at the poles much lower than that required for snowfall, to a temperature at the equator much higher than that at which snow can exist. So there must be some region between the two where snow ceases to fall; and to the pole-

(160) Fig. 44. — Snow Crystals.

ward of this region the length of the season during which snow falls increases towards the poles, while on the equatorial side snow never falls.

The equatorial limit of snowfall is in general on the continents at about the Tropics of Cancer and Capricorn, or $23\frac{1}{2}°$ of latitude; but in western South America it reaches much nearer to the equator. Over the oceans the latitude of 35° is about the limit.

It is found that in very cold weather snow does not fall; and as there must be a limit to the height of the temperature at which snowflakes can exist, so there must be some intermediate temperature of maximum snowfall. The subject has not been very extensively investigated, but some observations in the mountains of Germany have shown that, there at least, the snows occurred between the limits of 48° and 8° F. for the lower air temperature, with an average temperature of 30° F.

Measurements of Snowfall are first made by means of a graduated rod, with which the depth of the snow is measured at a number of places which are protected from the wind, and yet freely accessible to the snow. The snow in and over the rain gauge must then be melted, and the amount of the resulting water determined by the usual method of measuring rainfall.

The average density of recently fallen snow may be taken as about 0.10 (or 10 inches of snow make 1 inch of water); but snow is lighter (less dense) when it first falls than after it has stood for a while, lighter in cold than in warm weather, lighter in woods than on an open plain, lighter in gentle than in strong wind, and the upper layers of fallen snow are lighter than the lower. The density of snow may vary from 0.02 to 0.9 (the density of ice), the density of water being taken as 1.00; the conditions causing this great variation being the humidity of the air, the temperature, the wind velocity, the immediate locality whence the snow is taken (whether from drifts or protected places), the time that the snow has lain on the ground, the form of the snowflakes, the layer of snow chosen, etc. In general, in a cold

climate it may be assumed that the fallen snow is lightest in the early winter (in Siberia, 0.15), moderately dense in midwinter (in Siberia, 0.2), and most dense in spring (in Siberia, 0.3).

Dew. — Just as soon as the temperature of the surface of the ground (or of any other surface) falls below the dew-point of the adjacent air, the latter gives up part of its vapor in the form of small water drops, or dewdrops as they are called, which are deposited on the cooled surface. When the temperature of the surface lies below the freezing point of water (0° C. or 32° F.), then the dew is deposited in small ice crystals, and it is called *hoarfrost*. On account of the rapid cooling by radiation, especially on clear nights, the temperature of the ground and other solid substances becomes colder than that of the air above, and the dew-point and even frost point are reached by the ground and the adjacent air layer, while the air at the height of a few feet above the ground has a temperature several degrees above those points.

It has been supposed that in very dry rainless regions plants depend in a great measure on the dew deposit for their supply of water, but the amount of moisture obtained in this way is quite small.

In central Europe it was found that scarcely one inch depth of dew was deposited during a year, which is about 3% of the rainfall.

Night Frosts occur when plants are exposed to temperatures under the freezing point, and the conditions are much the same as for hoarfrost formation. Since by the condensation of vapor as dew, latent heat is given out and thus retards the cooling, the temperature during the dew-fall does not go much below the dew-point; so that when the dew-point lies above the freezing point, frost is hardly to be expected; but when the dew-point lies below the freezing point, frost may be expected on clear nights. The drier the air, the more likely will be the dew-point to be

below the freezing point, and therefore the more likelihood of a frost. This is seen on a large scale in the fact that frosts occur more frequently at inland than at coast stations.

Evaporation of Water.—A portion of the water which is found on the earth's surface is given off to the overlying air by the process of evaporation. The rapidity with which this evaporation takes place varies not only with the surface itself, whether it be a free water surface or wet ground or vegetable growth, but also with the temperature, the relative amount of water already in the air, the motion of the air, and the atmospheric pressure. In any case the amount of water evaporated varies directly with the amount of surface from which the evaporation takes place.

There is great variability as to the rapidity of evaporation from the ground; but it may be said that the smaller the size of the particles of earth, the more rapid will be the evaporation, since the capacity of the ground for holding water, and the capillary conduction of the water to the surface, increase with the fineness of the earth particles. From a hard-packed clay ground surface, almost no evaporation takes place.

The amount of evaporation from plants is enormously great. The passage of water through a plant from the ground to the free air is called *transpiration*. Five times as much water has been transpired and evaporated from a plant as from a water surface, and more than twelve times as much as from an ordinary land surface during the same time.

A short series of experiments has shown, that, according to the observed temperatures of the dew-point and the surface of the snow, the evaporation from the snow ex-

ceeds in amount the condensation from the air by means of its cold surface.

As a standard for measuring evaporation, the amount evaporated from a freely exposed but shaded surface of pure water is chosen; and the amount of water lost by evaporation is calculated by first measuring or weighing the whole amount when first exposed, and then that which remains in the vessel. Pure water evaporates probably 25 % more rapidly than salt water.

FIG. 45. — PICHE EVAPORIMETER.

A very convenient instrument for measuring the amount of evaporation during the warm season is the Piche evaporimeter, and it can be mounted beside a thermometer and read with as much ease as one. This instrument (shown in Fig. 45) consists of a graduated glass tube closed at one end. The tube is filled with water; and over the open end is placed a circular disk of porous paper, with a very small opening at the center. The instrument is mounted with the porous paper downward, and this last always presents to the air a wet surface from which the water is evaporated. As evaporation goes on, the upper part of the tube becomes emptied of the water, and the air makes its way up through the water column to fill the vacated space. The variation of the height of the water in the tube, as read off from the graduated scale, shows the amount of water evaporated from the tube. Since, however, disks of various degrees of porosity offer different resistances to the passage of the water, the instrument must be compared with a standard evaporimeter, or a disk of known porosity must be used.

Water when evaporated becomes a vapor which tends to distribute itself through the atmosphere somewhat after the manner of the air in an empty space. Vapor-laden

air is also transferred by the atmospheric currents to regions where there is a less amount of vapor.

Periodic Variations in Evaporation. — The amount of evaporation has a daily and annual march following that of the temperature. The evaporation is least at night, and greatest shortly after noon. The amplitude of daily variation is greatest in the interior of continents, and least on the coasts. The evaporation during the day is several times greater than during the night. During the year the least evaporation is in midwinter, and the greatest in midsummer.

At Nukuss (continental exposure in central Euro-Asia), the evaporation in January was about 1 inch, and in July nearly 12 inches. At St. Petersburg (seacoast), however, the evaporation was less than 0.2 of an inch in January, and less than 2.5 inches in July.

The amplitude of annual variation increases with the latitude and towards the interior of continents.

The Average Annual Amount of Evaporation varies greatly, but in general it decreases with the latitude. In the tropics, from a water surface, it amounts to perhaps 90 inches; in latitude 40°, to perhaps 30 inches; and in polar latitudes, to 10 inches. In hot desert regions, as in southwestern United States, it reaches even 150 inches.

The amount of evaporation possible over the land is greater than the amount of rainfall in most regions of the earth; but this amount of water does not evaporate from the ground, because there is not always sufficient moisture at the surface to keep up the rate of evaporation which could take place.

CHAPTER VII.

ATMOSPHERIC OPTICS AND ELECTRICITY.

Luminous Atmospheric Phenomena. — There are a number of luminous atmospheric phenomena which come to our notice, and which it will be convenient to mention at this stage. They are principally of two classes. The one pertains to the optical effects produced by and in the atmosphere; the other concerns the electrical condition of the atmosphere.

Transparency of the Air. — Any substance that one can see through is said to be transparent. But some substances are more transparent than others; that is, we can see through greater thicknesses of some substances than of others, and with greater distinctness.

We call air a transparent substance, but the degree of transparency is not constant. It varies with the density of the air and the amount and form of moisture which it contains. Thus at sea level, and for warm moist air, the transparency is least, while on high mountains with cool dry air the transparency is greatest.

Atmospheric Optics. — Light, in passing through the air, is subject to reflection, refraction, diffraction, and absorption; and, as a result of these, there arise a number of phenomena which must be briefly mentioned.

Refraction of Light is the bending of light rays. Light, in passing from a rarer to a denser medium, is bent from its straight course so that the angle which the ray of light

makes with the perpendicular to the boundary surface is less in the denser medium than in the rarer medium; and where it passes from a denser to a rarer medium, the angle which the ray makes with the perpendicular is increased. This angle in the first medium is called the *angle of incidence*, and in the second the *angle of refraction.*

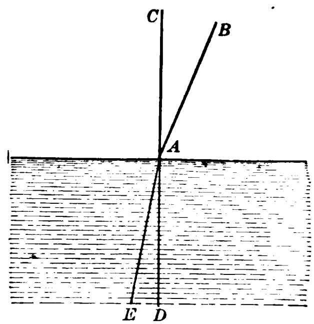


FIG. 46.

In Fig. 46, *BA* is the ray of light passing through a rare medium (as, for instance, air); and upon its entrance into a denser medium (as, for instance, water) the ray will be deflected from the direction of its path *BA*, and will take the course *AE*. If the line *CD* is perpendicular to the dividing surface between the two media, then *BAC* is the angle of incidence, and *DAE* is the angle of refraction.

The *amount of refraction*, or the difference between the angles of incidence and refraction, varies with changes in the density of the transparent media; and in the case of light entering the atmosphere from without, the rays are gradually bent more and more as the denser air is encountered.

The accompanying figure (Fig. 47) illustrates the bending of the solar rays entering the atmosphere. When the sun is below the horizon, at *C*, it would be invisible at *A*, on account of the curvature of the earth, if there were no atmosphere; but the solar rays entering the atmosphere near the point *B* are refracted so that they reach *A*, and the sun appears to be at *D*, though really at *C* below the horizon, either in the morning or in the evening. So that, in the polar regions, the sun is visible while it is in reality below the horizon, and is thus seen earlier and later during the time of polar sunlight.

The effect of atmospheric refraction is to slightly increase the amount of solar rays reaching a place.

Where the light enters the atmosphere at a small angular distance from the zenith, the amount of refraction can be quite accurately computed, but near the horizon it is very difficult to determine. For a zenith distance of 87.5° (i.e., 2.5° above the horizon), the refraction is computed to be 18.5′; but at the horizon it is about 30′.

The rays of light coming from the sun possess different wave lengths of vibration; and, as these rays pass from

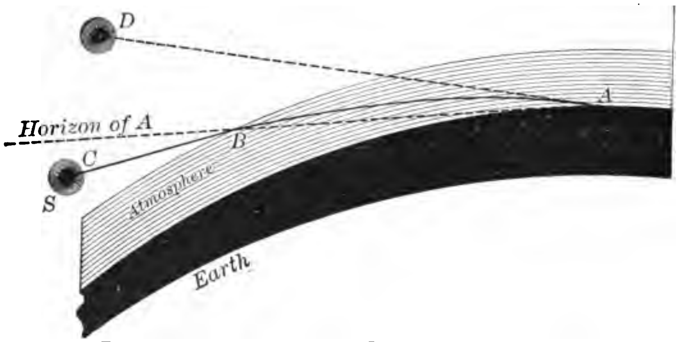


FIG. 47. — REFRACTION OF SOLAR RAYS BY THE ATMOSPHERE.

one medium to another of different density, there is more or less of a dispersion of the rays, and those possessing the longest wave lengths are refracted the least, and those with the shortest wave lengths are refracted the most; so that the rays which enter the second medium as white light are spread out by the refraction, and separated into the prismatic band of colors, — red, orange, yellow, green, blue, indigo, and violet, — ranged according to the wave lengths and refrangibility of the rays. Where any one color is observed, it shows that the rays with vibrations corresponding to this color are in excess of any others which may be present.

Reflection of Light is the throwing-back of light rays from the surface on which they fall. When the rays of light

strike a surface, they are reflected from this surface again in such a direction that they make the same angle with the perpendicular to the surface before and after reflection.

Thus, if a ray of light passes from *A* towards *B* (Fig. 48), it will be reflected from *B* towards *C*; and the angle *ABF*, called the angle of incidence, equals the angle *CBF*, called the angle of reflection. The reflection of light in the atmosphere is very important in its effects.

Reflection of Light from Dust Particles. — The reflection of light from the minute particles of dust in the atmosphere diffuses and scatters the light received from the sun in all directions, and illumines the sky. This illumination of the sky is greatest at low altitudes, where the air contains more and larger dust particles than at higher altitudes.

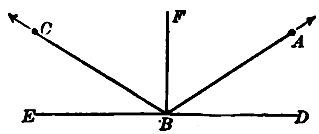


FIG. 48.

Mirage. — When one layer of air lies in contact with another layer of different temperature and density, the boundary surface between the two, if it is sharply defined, will reflect light quite perfectly. When the observer is above such a surface, he sees objects beyond and above (but not below) this surface reflected from it as from a horizontal mirror. In case the observer is below this boundary surface, he sees the objects beyond and below (but not above) reflected from it. The images are inverted. This phenomenon is called *mirage*.

Diffraction of Light is the dispersion or breaking-up of light rays. When a beam of light passes through an aggregation of exceedingly minute particles, or when it strikes a surface roughened or divided into very small surfaces, the rays are broken up and scattered or diffracted. The interference of some of these rays causes

some of the color rays to partially, and others to wholly, disappear.

Colors of the Sky.—The light rays having the more minute wave length (those nearer the violet end of the spectrum) are diffracted by the more minute as well as by the coarser particles, while the rays of greater wave length (those nearer the red end of the spectrum) are diffracted more by the coarser particles; and it is on this "selective" diffraction and reflection that much of the coloring of the sky and other more definite optical phenomena depend. The blue color of the sky (away from the sun) is due to the more powerful diffraction of the blue rays, so that rays of that color are reflected towards the observer, while those of coarser wave length pass on and do not reach him. As the observer turns his eyes gradually towards the sun, the sky loses its blue color, and takes on a more neutral or white color, because the direct rays which now reach him do not contain to any great degree the blue component which had been previously reflected toward him from other directions.

When the sunlight passes through the atmosphere, we lose more and more of the colors having smaller wave length, the thicker and denser the air mass through which the rays come, so that the sun's rays reaching us contain the most blue when in the zenith; but as the sun approaches the horizon, the rays at the blue end of the spectrum become scattered, and lost in the air, and we receive only those at the red end, and consequently the sun appears red at the horizon. If the sun could be viewed without the intervention of an atmosphere, it would present the distinctly blue color.

The glow of the sky which accompanies the rising and setting of the sun is due to the diffraction and reflection of light by the minute particles in the air. As the sun

sets below the horizon, the intervening portion of the earth cuts off the rays from the lower atmosphere (at the point of the observer), and the diffracted rays forming the glow appear in the upper atmosphere only. When the particles of matter in the air are unusually numerous at high altitudes, the sunset glow of the sky is extraordinarily brilliant. Such a condition prevailed during 1883 and a few subsequent years.

Corona and Halo. — The sun or moon is occasionally surrounded by one or more well-marked rings or circles of light, but not always of the same diameter or color. These rings are divided into two classes, — the *corona*, of small diameter; and the *halo*, of greater extent. In the corona, the color of the inner part of the ring is blue, and of the outer, red. For the larger halo, the order is reversed; and red is on the inside and blue on the outside of the ring. The radius of the halo is about 22°, 45°, or 90°. The corona may vary from 1° to 10° or more in radius.

The corona is a diffraction phenomenon, while the halo is due to refraction and reflection.

The corona may consist of several rings concentric with the sun or moon, and occurs when mist or thin clouds partially obscure those luminaries. Coronæ are formed by the diffraction and interference of light caused by the small water particles in the cloud. The larger the water particles, the smaller will be the ring; and it is when the particles are of different sizes that coronal rings of different diameters exist at the same time.

The halo occurs in connection with the higher cirrus clouds only. Halos are more frequently observed than coronæ. The faint halos are much more easily detected by observing in a black mirror the reflection of the light center and its neighborhood. There is great variety in

the distribution of halos, — sometimes the rings are wholly separate, and sometimes they intersect. The points of intersection are unusually more or less bright patches, and are called ***mock suns***, or ***sun dogs*** (Fig. 49).

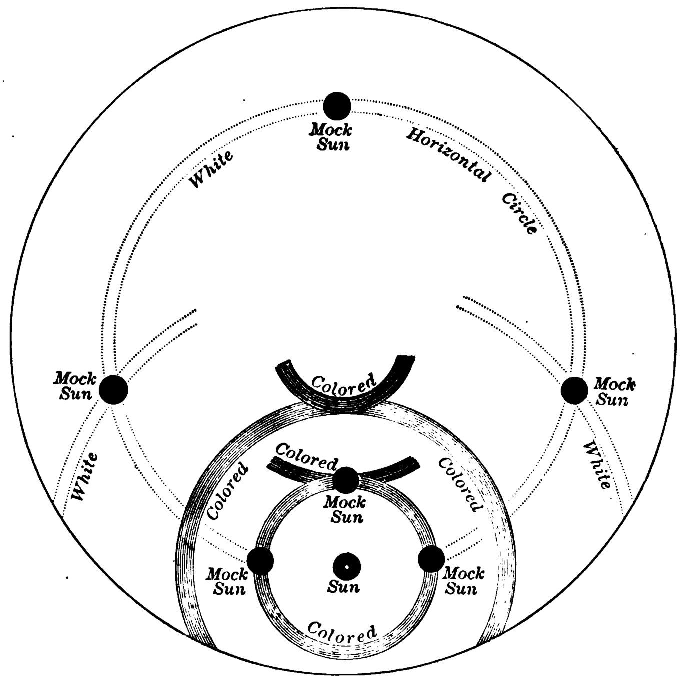


FIG. 49. — HALO PHENOMENA.

In general, the halos may be divided into the three following classes: —

1. Rings which have the sun at the center, and have a radius of about 22°. The rings are about 1° wide, and are colored with the red on the inner side, where they are sharply defined, but on the outer violet side they gradually vanish. The space directly around the sun is bright, but towards the ring this brightness gradually disappears.

2. Rings which pass through the sun and have a radius of about 45°.

These rings are colored, and have the red side towards the sun. The width of the ring is about 3°, and although not so intense as the first kind, yet the colors are more distinctly separated. Sometimes a ring of this nature is visible around the whole sky parallel to the horizon at the altitude of the sun.

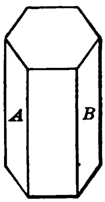


FIG. 50.

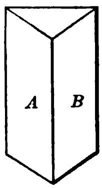


FIG. 51.

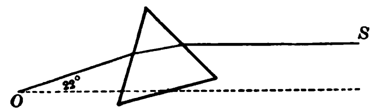


FIG. 52.

3. Rings which turn a convex side towards the sun, and have a radius of about 90°. These are of weak intensity and of indefinite color. Only portions of the ring usually are visible; and they touch some of the smaller rings which encircle the sun.

Explanation of Halos. — These ring phenomena are mainly due to the refraction of the light by the ice crystals or ice needles of which the cirrus clouds consist.

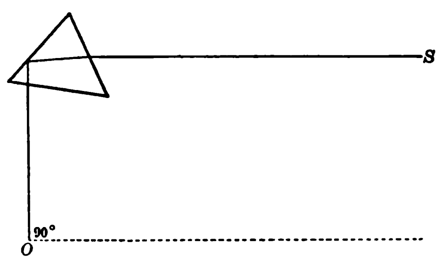


FIG. 53.

These needles are assumed to be six-sided (Fig. 50), and to hang suspended in the air. When they hang vertically, and the rays of light pass through as if the needles were triangular (Fig. 51), then the light is diffracted as shown in Fig. 52, and the ring has a radius of 22°. When some of the ice needles are inclined to the vertical, then the light may enter at another surface, and be diffracted through 90°, and the ring of 45° will result.

In order to cause the ring of 90° radius, the light must pass through the crystal, as shown in Fig. 53, and suffer total reflection.

The horizontal ring parallel to the horizon is due to the reflection of the light from the outer surface of the ice crystals, which are at the altitude of the sun, and therefore possess the proper inclination of the surfaces to reflect the sun's rays to the observer's eye; and since no refraction occurs by means of the ice needles, there is no coloring, and the ring appears white.

The broken rings or arcs are likewise reflection and not refraction phenomena.

The Glory, Brocken Specter, or Fog Image is analogous to the coronal phenomenon. It is brought about by the sun casting a shadow of the observer on a fog or cloud bank. This shadow, sometimes of huge dimensions, is surrounded by a glory of light, which is caused by the diffraction of the rays of light by the water particles (near the observer), and the resulting separation of the prismatic colors which are reflected to the eye of the observer by the fog particles.

Sometimes, where the observer is on the top of a mountain, the whole peak may have its shadow thrown against the fog bank. At times a seemingly distant and white fog bow encircles the shadow of the observer and the glory which surrounds the head. A glory of light is also sometimes to be seen around the shadow of the head of the observer when it falls on bedewed or wet grass.

Rainbows are produced by the refraction of the sun's rays by means of the raindrops in the air, after which the separated band of colors is then reflected to the observer's eye. The center of this bow is opposite to the sun. The radius of the bow is 40° to 42°, with sometimes a fainter secondary bow outside, with a radius of 50° to 54°. The inner bow has the blue color on the inside, and the red on the outside; while this order is reversed for the outer bow. Rainbows are most frequent in the local showers in which the sun suddenly breaks through the clouds at the edge of

the storm. Where there is a wide distribution of cloudiness, as in our long-continued rains, the proper conditions for the formation of the bow will seldom be present. Rainbows are visible when the sun is at low altitudes; and the nearer the latter is to the horizon, the greater will be the length of the bow visible in the sky. The bow is 180° in length, or a semicircle, when the sun is close to the horizon.

Atmospheric Electricity. — We find the atmospheric air usually in a state of positive electrification; that is, it is generally charged with positive electricity, which is the condition of glass when rubbed with silk. Very great fluctuations, however, occur, especially during thunderstorms, snowstorms, etc.; and the electrification becomes sometimes negative, which is the condition of glass when rubbed with resin or sealing wax. The origin of atmospheric electricity is still unaccounted for, but it may be due to the frictional action of the air. The degree of electrification of the air is measured by means of an instrument called an *electrometer*.

There are two classes of atmospheric electric phenomena which we shall notice: they are *auroral* displays, which may partly occur in the lower atmosphere, and partly in the highest layers; and *lightning* displays, confined to the lower air layers.

The Aurora — called in the northern hemisphere the *aurora borealis*, or northern lights, and in the southern hemisphere *aurora australis*, or southern lights — is an illumination of the sky which occurs in middle and polar latitudes, and which has a zone of maximum frequency from the 60th to the 70th parallel of latitude north and south of the equator. It rarely occurs within the tropics. It is possible that some of the widespread, intensely active, auroral phenomena have an electromagnetic origin; while those of

local occurrence may be simply the electrostatic charge of the air layer, rendered luminous by atmospheric changes.

As usually observed by us, the aurora consists of an arch or band of light. It may, however, have various forms, such as that of an arch, ribbon, collection of beams, corona, and haze or diffused light. The auroral arch usually spans the sky in an east-westerly direction. In the middle latitudes it is seen to the north, and in the extreme northern latitudes to the south. The dimensions are variable. (A band observed in 1893 was supposed to be 15 miles wide, perhaps 250 miles high, and over 1,000 miles long.) In general, the arch may be said to vary between an unknown lower limit and 300 miles in altitude. The arch sometimes varies in intensity in different parts. Frequently band-like streamers of light are seen perpendicular to the arch, and they extend to an unknown height. Sometimes these streamers are constant during their existence, and sometimes they are intermittent and pulsating, flashing out at intervals.

Auroræ in the northern hemisphere are least frequent in January and June, and most frequent in March and October. There seems also to be an 11-year period of frequency such as was found by astronomers for sun spots.

Lightning. — The air between two clouds charged with electricity, or between a charged cloud and the earth, is subject to an electrical strain or pull. When this strain upon the air column becomes too great for further resistance, a disruptive discharge takes place. This discharge may vary in character from the invisible silent lightning to the violent impulsive rush discharge which has an enormous amount of energy. The lightning flash is the luminously heated air in the path of the discharge. The flash may have a duration varying from $\frac{1}{300}$ of a second to a second. The thunder is but the crackle of the electric

discharge, but its sound is frequently reënforced and prolonged by numerous refractions and reflections from adjacent clouds and other objects.

The needle of an electrometer which indicates the amount of the electrical strain will show by its fluctuations or "breaks" not only all of the flashes or discharges of lightning which are visible, but also the silent discharges which are of frequent occurrence but are invisible.

The typical forms of lightning flashes (which have been studied by the aid of photography) are, —

1. *Stream lightning*, which consists of a plain, broad, smooth streak or flash of light.

2. *Sinuous lightning*, which consists of a flash following some one general direction; but the line is sinuous, bending from side to side. This is the most common type.

3. *Ramified lightning*, in which part of the flash appears to branch off from the main stem like the branches of a tree from the trunk; but whether these branches issue from the trunk or unite with it, is unknown.

4. *Meandering lightning*, in which the flash appears to wander about without any definite course and form irregular loops.

5. *Beaded lightning*, in which a series of bright beads of light appear along the white streak of lightning.

6. *Dark flashes* of lightning have also been photographed.

The causes of the different forms are not well understood.

Since the electrical discharge between the clouds and the ground will take place more readily where the distance is least (other conditions being equal), it is quite probable that the points of the clouds which appear suspended from the main thundercloud mass will the most frequently form the places from which the discharge takes place. In case of a tornado-funnel cloud, in which the cloud reaches nearly or quite to the earth, the condition for a discharge along this would be most favorable. Also at the time of the greatest rainfall the electric discharge would be greatly facilitated by the excellence of the conduction which would

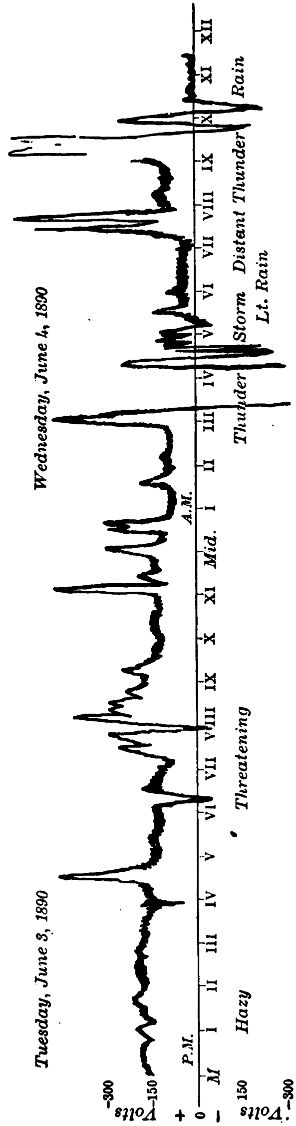


Fig. 54. — Atmospheric Electric Potential (after McAdie).

be established by the falling water between the cloud and the earth.

Periodic Changes in Atmospheric Electricity. — There is a periodic daily change in the degree of electrification of the air. There are times of maxima at about eight or nine o'clock in the morning and evening, and of minima at two to four o'clock at night and in the afternoon. The cause of this periodicity is not known with certainty.

There is an irregular annual periodic change, of which we can only say that in general the maximum electrification occurs in the cold season, and the minimum in the warm season.

The accompanying diagram (Fig. 54) shows the fluctuations in the potential of atmospheric electricity during a couple of days, as observed by means of an electrometer. In the curve not only do we find rapid fluctuations in the positive electrification, but there are also sudden changes from a powerful positive to an almost equally powerful negative electrification when the thunderstorm is in progress.

St. Elmo's Fire is an electrical phenomenon sometimes

seen as a continuous luminous discharge from elevated points. It usually has the appearance of jets of flame issuing from the points of the objects, and is accompanied by a hissing or buzzing noise. It is most frequent in winter, and is almost invariably accompanied by a heavy fall of soft hail or snow.

CHAPTER VIII.

GENERAL CIRCULATION OF THE ATMOSPHERE.

Some Preliminary Ideas concerning Air Motions. — We come now to the study of the systems of movements of the air, which have already been mentioned in the chapter on winds. It is first necessary, however, to state in a general way the nature of air motions, and the changes which air undergoes, and its conditions with respect to the surrounding air, when subjected to these motions, especially to vertical motions. These are briefly stated in a necessarily disconnected manner, and their bearing or importance cannot be fully seen until they are applied to proper cases as they come up in their natural order.

General Air Motions. — Atmospheric motions may in general be divided, according to direction, into two classes, — *horizontal* and *vertical.* Air currents are seldom inclined at a large angle (such as 20° or 45°) to the horizontal or vertical directions.

Horizontal air motions are those nearly parallel to the earth's surface, and are produced by gradient forces effective in those directions. In horizontal currents the changes in density and temperature of the moved air mass are usually very gradual and relatively slight.

Vertical air motions arise from two causes. They are either a natural consequence of the horizontal motions, in order to preserve a continuity of the air taking part in these motions when they have dissimilar directions; or they are

due to the endeavor of a local mass of air having an abnormal condition to move to a place where its condition will be normal, i.e., the same as that of the mass of surrounding air. In vertical currents the changes in density and temperature of the moved air are rapid and of relatively great magnitude.

Centrifugal Force. — If a body (a mass of air, for instance) has a motion in any direction, it will continue to move in a straight line with constant velocity, unless it is acted upon by some outside force or is subjected to some sort of resistance. When a body is constrained to move in a circular path, it still has a tendency to move in a straight line, which would be in the direction of the tangent to its circular path, at whatever point the body happens to be. The force with which the body presses outward, in its endeavors to depart from the circular motion, is called the *centrifugal force*. This force would be the pull exerted on a cord by a weight attached to one end of the cord, if it were whirled around in a circle of which the other end of the cord (or any part of it) was the fixed center of revolution.

In case a mass of free air is forced to move in a curvilinear path, it will exert a certain amount of centrifugal force outward from the center of rotation; and the amount of this force is exerted as a pressure against outer adjacent masses of air. Since air is an easily yielding and compressible substance, the effect of this centrifugal force is to diminish the amount of air within the circle, and to increase the amount without the circle. Where the whole of a mass of air has a rotation around a point at its center, the centrifugal force causes the air to recede most at this center, and then less and less at increasing distances from the center.

Conservation of Areas. — When a body (air, for example) is acted on by a central force, or the opposite, and has a gyratory motion around this center, then the varying line connecting this body with the center (called *radius vector*) sweeps over equal areas of space in equal times. Therefore when the air approaches the center of revolution, its velocity of revolution must increase as the radius vector shortens; and conversely, when it recedes from the center, the velocities decrease.

Energy of the Air. — While, from the fact that we live in the lower part of the atmosphere, we must deal mostly with the winds near the surface of the ground, yet it must not be assumed that the upper half of the air (that is, the portion above the altitude of about three miles) can be neglected in studying its conditions, for it has been computed that in this upper half of the atmosphere there is six times as much gradient force or energy productive of motion as in the lower half. The altitude of about 40,000 feet above sea level, which is about the altitude of the highest clouds (the cirrus clouds), divides the energy of the atmosphere into two parts. The amount of gradient force or energy above this altitude is equal to that below it.

The Heat in the Air. — The heat which is in the atmosphere is, as we have seen, derived by a number of processes. There is first absorbed about half of the direct solar heat; then about half of the heat reflected from the ground and water surface of the earth; then all the heat supplied from the earth's surface by convective currents, and the heat supplied by radiation from the earth's surface (just as it is radiated from the sun). All this heat must be lost from the air through a process of radiation from the air itself, which must not be confused with the radiation through the air from other objects.

Radiation of Heat from Air. — The process of radiation of heat from, and of absorption of heat by, layers of air within the atmosphere, depends on the difference of temperature, and perhaps on the difference of density, of these layers; and without doubt a considerable portion of the heat which escapes from lower air layers passes directly through the upper air into space, and the portion of heat which is absorbed by the air passes in its turn through the layers still above, and finally makes its escape.

This radiation of heat from the air itself is shown by the fact that the cooling of the air which takes place on a clear night is quite general; while, if it depended on the conduction of the heat from the air to the cooler ground to replace that lost by radiation from the latter, there would be a cooling of the air to a height of not more than 10 feet during, say, twelve hours of darkness.

The rate of actual cooling of the air by radiation varies approximately as the difference in temperature of the air and its surroundings, and it is therefore probably least at the middle altitudes of the atmosphere.

Comparison of Solar Heat directly absorbed by, and radiated from, the Air. — The amounts of heat absorbed by the air and radiated from it again are nearly equal, under the maximum conditions of absorption, during the same period of time. But since the radiation from the air into space goes on at all times, and since the direct absorption of the solar heat rays can take place only when the sun is shining, the outward radiation is much more effective in cooling the air than is the direct action of the sun's rays in warming it.

Adiabatic Heating and Cooling of Air Dynamically. — The adiabatic heating and cooling of masses of air by the variation of the air pressure, due to the ascending or

descending movement of the air, is one of the most important factors in atmospheric events.

Cooling of Dry Air in Ascending, and Heating in Descending Currents. — Theoretically, the rate of cooling in ascending and the rate of warming in descending dry air is 1° F. for each 183 feet change in altitude, whatever the temperature of the air. If the air is in any way subjected to a change of pressure equivalent to that which would occur in passing through 183 feet in altitude, it will experience the same changes of temperature as it would if actually moved through this distance.

Action of Ascending Currents of Moist Air. — We have just been considering the air when it is dry; but when moisture is present, as it always is in the atmosphere, we have no longer the simple conditions just mentioned. The air, as it cools adiabatically in its ascent (as just discussed), has a diminished capacity for moisture; and when the limiting temperature is reached, at which condensation of the moisture present occurs, then, with the condensation which follows, there is a freeing of latent heat, which retards the further cooling of the air. Thus there is heat added to the air mass; and if the air keeps on in its ascent, the decrease of 1° F. per 183 feet of ascent is diminished by the amount of latent heat supplied by this condensation. But until condensation does actually take place, the rate of this decrease of temperature is nearly 1° F. per 183 feet, even when moisture is in the air.

Descending Currents of Moist Air. — The rate at which air becomes heated in descending is 1° F. for each 183 feet, whether the air is dry or saturated (except where rain is held in the air, or where supersaturation exists). Thus the rate of change of temperature may be very different in ascending and descending currents of moist air.

The Actual Rate of Decrease of Temperature with the altitude is less than that assigned by theory, and varies somewhat on account of various irregularities in the existing atmospheric conditions. It is greater in summer than in winter, and differs in cloudy weather from that in clear weather. Still these irregularities are but departures from the laws just mentioned, and their causes would have to be examined separately.

Indifferent Equilibrium. — If the temperature of the unsaturated quiescent or horizontally moving air adjacent to or surrounding a vertical current of air decreases 1° F. for each 183 feet increase in altitude, then the air is said to be in *indifferent equilibrium.* For unsaturated air which is moved upward or downward, and which changes its temperature at this rate, will remain in the new position in which it is placed, because its change of temperature has been just such as to accommodate itself to the temperature of the surrounding air in its new position.

Stable Equilibrium. — Unsaturated quiescent air which decreases in temperature at a rate less than 1° F. for each 183 feet of increase in elevation is said to be in *stable equilibrium:* for when a portion of such air receives an upward motion, its density gradually becomes greater than that of the surrounding air at the same level, and it would sink back again to its starting place if the force which moved it ceased to act.

The differences in air density depend on the relation of decrease of temperature with altitude, in the air which is in motion, to that in the air which surrounds it. Since the rate of decrease of temperature of the unsaturated ascending air is about 1° F. for each 183 feet of ascent, then, if the temperature of the surrounding quiet air decreases at a rate less than this, the ascending air will cool the more rapidly, and will become the denser as it ascends; and when the force which made it

ascend is spent, the air which has ascended will fall back again to its original position. Similarly, if a downward current of air becomes heated faster than the surrounding air, then it becomes lighter than the air at any lower altitude at which it may arrive; and when the force which causes it to descend is spent, the air will rise again to its former position.

Unstable Equilibrium. — Unsaturated quiescent air which decreases in temperature at a rate greater than 1° F. for each 183 feet of increase in elevation is said to be in *unstable equilibrium:* for when a portion of such air is moved upwards it becomes lighter than the surrounding air, and when moved downwards it becomes still heavier than the surrounding air, at any level which it may reach; and thus there is a tendency for the moving air to continue in the direction in which it is started.

This condition occurs when the temperature of the moving air varies at a rate less than that of the surrounding quiescent air.

Conditions of Equilibrium illustrated numerically. — In the accompanying table (from Ferrel's "Winds"), the first column gives the altitudes; the second column, an assumed temperature at these altitudes which would produce stable equilibrium; the third column, the temperatures which would produce indifferent equilibrium; and the fourth column, an assumed temperature which would produce unstable equilibrium in case vertical currents arise.

Altitude.	Temperatures for Stable Equilibrium.	Temperatures for Indif. Equilibrium.	Temperatures for Unstable Equilibrium.
Feet.	F.	F.	F.
9,840	42.8°	32°	21.2°
8,200	50.0	41	32.0
6,560	57.2	50	42.8
4,920	64.4	59	53.6
3,280	71.6	68	64.4
1,640	78.8	77	75.2
000	86.0	86	86.0

Here we see, that since a vertical air current has a change of temperature with the altitude of about 1° F. per 183 feet, then, when the rate of change of temperature in the surrounding quiescent air is less than this, the condition is that of stable equilibrium; when it is equal to it, the condition is that of indifferent equilibrium; and when it is greater, the condition is that of unstable equilibrium.

General Circulation of the Atmosphere. — The winds belonging to the general circulation of the atmosphere are the hemispherical systems of winds extending from the equator to the poles. It has been found by observations of wind vanes, that in the *northern hemisphere*, in the lower air, there is a decided prevailing wind blowing from the east or northeast over the region extending from near the equator almost to the tropics; while from a little beyond the tropics to the polar regions there is a prevailing wind from the west or southwest. Directly at the equator and in the region of the tropics, there is no such prevailing direction of wind, and in fact these may be designated regions of relative calms. These surface wind directions are shown in a general way by the arrows (drawn flying with the wind) on the charts, Figs. 26 and 27, and are also shown by the heavy arrows on the inner, shaded portion of Fig. 55.

At high altitudes, mostly in the regions above the highest clouds, there are prevailing winds blowing from the west or southwest over the entire hemisphere from the equatorial to the polar regions.

Between the winds of the upper and lower regions, in which the wind direction is mostly poleward, and extending to the ground in the polar regions, there exist return currents flowing equatorward, which prevent the accumulation of air in the polar regions.

It must not be supposed that these general wind directions are continuous and uninterrupted. As a matter of fact, there are continually occurring within them local air

movements of sometimes thousands of miles lateral extent, which entirely change, during their continuance, the direction of the main air currents. So it is only by observing the winds for a great length of time that the average normal direction can be determined.

In the *southern hemisphere* these wind directions are somewhat reversed: the prevailing lower currents between the Tropic of Capricorn and the equator are from the

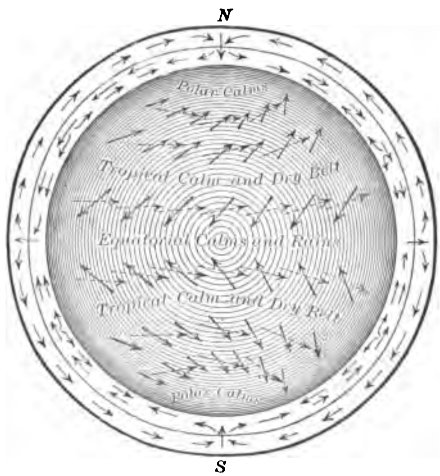


FIG. 55. — DIRECTIONS OF PRIMARY AIR CURRENTS (AFTER FERREL).

southeast, and those to the poleward of the tropic are from the northwest.

The upper air has a general movement from the west and northwest, while for middle altitudes the movement is from the pole equatorward. In the polar regions this counter or return current reaches to the ground.

Schematic Diagram of the General Atmospheric Circulation. — The accompanying diagram (Fig. 55) shows the main features of the general atmospheric circulation. On

the shaded sphere, the complete arrows show the direction of the lower air currents, and the broken arrows that of the upper currents; the arrows flying with the wind.

The outer, unshaded portion of the figure shows the circulation projected into the plane of the meridian; and the position of an isobaric surface at low altitudes is represented by the inner, and that of an isobaric surface at high altitudes by the outer, encircling line.

On the body of the sphere we see the eastward upper current at all latitudes, and the lower current eastward in the higher latitudes, but westward in low latitudes.

The causes of these wind movements, and the connection between them, have been studied out with great care; and some of their details are now to be entered upon, after comprehending which, the general air motions shown in Fig. 55 can be more readily understood.

The Cause of the General Circulation of the Atmosphere is primarily the large but not constant difference in temperature existing between the equatorial and polar regions. These atmospheric motions embrace, generally speaking, a whole hemisphere each side of the equator. The difference in the air temperature between the poles and the equator amounts (at a few feet above the ground) to about 80° F. for the average of the entire year. Since the colder the air the denser it is, any isobaric surface is nearer the ground at the cold pole than it is at the warm equator. There exists, then, a declination of isobaric surfaces and barometric gradients in the direction of the poles, and there is a movement of the air down these isobaric surfaces after the manner already described. This movement down the decline of the isobaric surfaces from the equator towards the pole continues until the altitude of these isobaric surfaces becomes so lessened at the

equator and increased at the pole as to bring them to the same level throughout. This motion originally occurs at all altitudes where the isobaric surfaces decline towards the pole; but it increases with the altitude, because the difference between any two upper air layers must be added to the differences which already existed for the layers below. In Fig. 56, *D* is the equator, *CC* are the poles, and *BAB* an isobaric surface. The arrows show the direction of the flow of air.

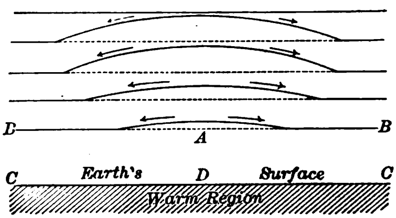


FIG. 56. — MERIDIONAL AIR MOTION.

As a result of these air currents towards the poles, there arise (as is explained later) other nearly horizontal currents towards the equator, and upward and downward currents connecting them (see Fig. 57). It has been computed that if the average temperature of an air layer 33,000 feet in thickness, at latitude 30°, were 40° F., and if the temperature at the pole were 90° F. less than this, then the upper isobaric surface of this layer would be depressed so much that it would be nearly 6,000 feet lower at the pole than at latitude 30°; and a mass of air gliding down this inclined surface would attain at the pole a velocity of over 600 feet per second, if it started at latitude 30° from a position of rest, and its motion were frictionless and undisturbed, and under the influence of gravity alone.

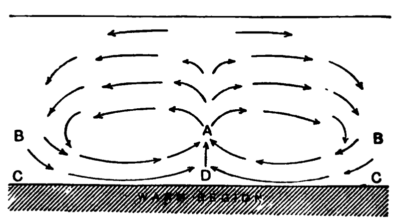


FIG. 57. — IDEAL MERIDIONAL AIR CIRCULATION.

At the earth's surface the observed greatest pressure is at about latitude 30°, and the calculated depression of the isobaric surface is 260 feet at the equator, and in the southern hemisphere at the pole it is about 650 feet; while in the northern hemisphere the maximum de-

pression is only about 325 feet, and is at latitude 60°; from thence to the north pole the depression lessens.

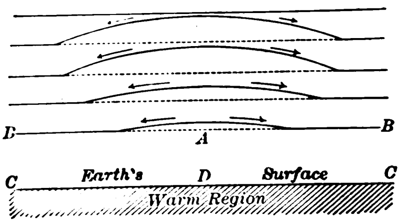
On the basis of observed air pressures at sea level, it has been computed that at an altitude of 33,000 feet the depression of the isobaric surface amounts to over 2,800 feet at the north pole, and 3,150 feet at the south pole.

Diurnal Rotation of the Earth. — In addition to the barometric gradient due to the difference in the temperature at the equator and the pole, there is another influence powerfully affecting the general air motions, and this is the diurnal rotation of the earth on its axis. In considering the matter of air motions, however, it is best first to examine the conditions resulting primarily from the differences of temperature alone, and then to add the modifying influences of the earth's rotation.

Ideal General Circulation of the Air without Rotation of the Earth. — If the motion of rotation of the earth on its axis is not considered, then the movement of a particle of air is either north and south, or upward and downward, in the plane of its meridian.

Starting with the high temperature at the equator, and the low temperature at the pole, and the resulting inclination of the isobaric surfaces from the equator towards the pole, we find the equatorial air flowing poleward down these surfaces in the direction of the lowest level with a velocity which increases with the altitude. The upper air, then, flows poleward most rapidly; but at the earth's surface there is no motion, because at first there is no gradient at this lowest surface, the *amount* of air above being everywhere the same. Just as soon, however, as any of the upper air leaves its position in lower latitudes, there is a lessening of the actual weight of the air (pressure at the surface of the earth) there, and an increase

equator and increased at the pole as to bring them to the same level throughout. This motion originally occurs at all altitudes where the isobaric surfaces decline towards the pole; but it increases with the altitude, because the difference between any two upper air layers must be added to the differences which already existed for the layers below. In Fig. 56, D is the equator, CC are the poles, and BAB an isobaric surface. The arrows show the direction of the flow of air.



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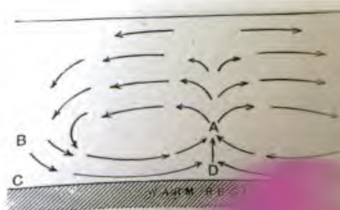

FIG. 57. — IDEAL MERIDIO

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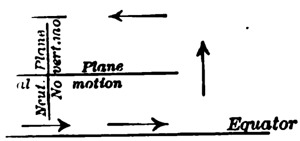

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towards the pole, by reason of the air which has flowed thence; and a countercurrent of air sets in along or near the surface of the earth from the region of the pole towards the equator, to take the place of the air which moves away above from the latter region. We thus have a current of air above flowing poleward, and a current below flowing equatorward; and in order to satisfy the conditions of continuity, which allows no gaps to occur in flowing fluids, there must exist vertical currents at the ends of these two horizontal currents, in order to complete the circuit by connecting them.

The direction of the upward and downward connecting currents is conditioned by the direction of flow of the main horizontal currents, for they must be such as to keep up a connected flow of air. Thus an upward current at the equator and a downward current at the pole are needed to meet this requirement. The circuit is shown in the diagram (Fig. 57).

Since the temperature differences at the poles and the equator are perpetual, and the force (gravity) which causes the poleward motion is a continuous one, this circuit of the air current is uninterrupted in action after it once begins. And, moreover, since the cause of the motion acts continuously, there would be a continually increasing velocity in this circuit, if it were not for the loss of motion through friction, and the interference or mixing of air masses having different directions and velocities of motion.

Regions of Separation between the Horizontal and Vertical Currents, respectively. — It is evident that there must be some region between the upper poleward air current and the lower current towards the equator, in which there is no horizontal motion one way or the other; and likewise between the upward and downward currents at the equator

and the pole there must be a region where there is no vertical motion, and where the motion changes from one direction to the other. These regions of no motion are called *neutral planes* (see Fig. 58).

While at the neutral plane between the upper and lower air currents there is no horizontal motion in either direction, yet, with increase in altitude above this plane there is a gradual increase in the poleward velocities, and below this plane there is an increase in the equatorward velocities down to within a few hundred feet of the ground, where the friction of the earth retards the air motion, and makes it decrease again with nearer approach to the

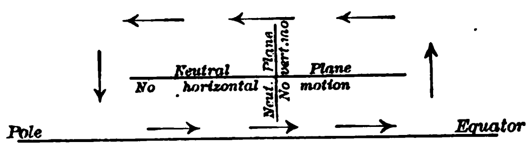


FIG. 58. — REGIONS OF NO AIR MOTIONS.

ground. Likewise at the neutral plane between the vertical air currents at the pole and the equator, there is an increase of the downward motion towards the pole, and an increase of the upward motion towards the equator. At the intersection of these neutral planes there is no motion in any direction, and consequently a calm.

Effect of the Convergence of Meridians. — We have been picturing the motion along the meridian. Now, it is necessary to take into account the fact that the meridians converge more and more as the equator is departed from, and meet in a point at the pole. The effect of this convergence is to gradually decrease (with the decreasing size of the parallels of latitude) the amount of space (between the different meridians) which can be occupied by the air above,

which flows poleward; and to gradually increase (with the increasing size of the parallels of latitude) the amount of available space which can be occupied by the air below, which is flowing from the polar regions equatorward. Since, with assumed constant velocity of motion, about half of the air must be to the poleward, and half to the equatorward, of the vertical neutral plane, the effect of the convergence of the meridians is to force the vertical neutral plane equatorward; and it must lie somewhere about latitude 30°, because this divides the surface of a hemisphere (and consequently the cubic space between the earth's surface and the outer limit of the air) into two equal parts. This affects both the velocities and paths of moving air masses.

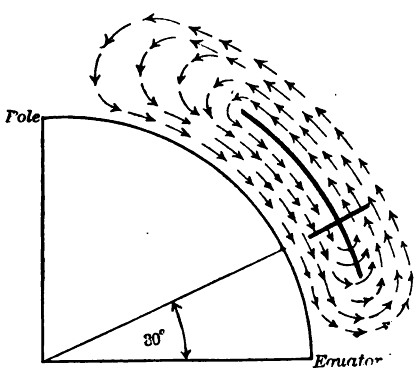


FIG. 59. — REGIONS OF AIR MOTIONS AND CALMS ALONG A MERIDIAN.

Meridional Movement of an Air Mass. — If we follow the motion of a mass of air starting out from a position near the equator but at a considerable altitude above the earth's surface, and above the level of the horizontal neutral plane (Fig. 59), we shall find that the horizontal motion of the air mass as it moves towards the pole is accelerated for a time until it reaches some intermediate latitude, and then it is retarded as it approaches its polar limit; and with this retardation there is a gradual downward fall of the air until it reaches the point where it is nearest the pole, which occurs at the altitude of the horizontal neutral plane.

The air mass now commences its return journey in the lower current towards the equator: there is at first a continued downward motion, and then it becomes more nearly horizontal and continues so until after the neutral vertical plane is passed; there is an acceleration of velocity until an intermediate latitude is reached, after which the horizontal motion is retarded, and there is a gradual rise of the air mass until it reaches its place of starting.

The circuit made by the air masses is somewhat elliptical (oval) in form; and it is seen that some of the air masses move in large orbits which extend from the equator almost to the pole, while others move in smaller orbits, some of which extend only a short distance each side of the intersection of the two neutral planes. All these motions are along the meridians.

The horizontal current reaches its greatest velocity in middle latitudes, and vanishes at the equator and the poles. The vertical current disappears at the earth's surface and at the outer limits of the atmosphere. The descending velocity near the pole is less than the ascending velocity near the equator, because the downward current meets with more resistance than the upward current. The downward-moving air meets air of increasing density, and the surface of the earth offers resistance to its motion. The upward current carries the air into masses of less density, and there is no resistance from a rigid body such as the earth's surface. The velocities of the vertical currents are relatively small, but their great meridional extent magnifies the importance of their action. The relation of the horizontal flow of air to the vertical is, roughly speaking, as the radius of the earth to the height of the atmosphere, because, while the greatest length of the horizontal current is from the equator to the pole, that of the vertical current is from the earth's surface to the upper limit of the atmosphere.

The Effect of the Earth's Rotation, on the air currents passing between the equator and the poles along meridians, and on the vertical air currents which connect these hori-

zontal currents, is a matter which has not yet been investigated in all of its details, and it is so difficult that only a meager outline of it can be presented here. There is one general theorem on which are based all modern explanations of this action, and this is as follows: —

If a free-moving particle (such as air) moves along near the earth's surface, there is a force arising from the diurnal rotation of the earth which deflects it to the right of its course in the northern hemisphere, and to the left of its course in the southern hemisphere. The amount of this force increases with the mass and the velocity of the particle, and also with the increase of latitude. It is zero at the equator, and greatest at the poles.[1]

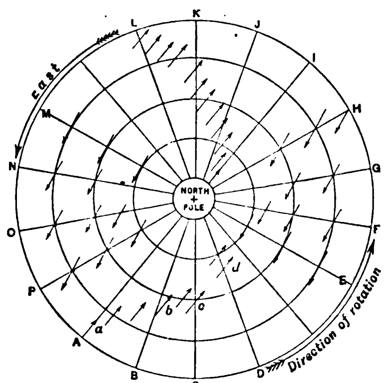


FIG. 60. — DEFLECTION OF THE WINDS BY EARTH'S ROTATION.

The following illustration will show the general nature of this deflecting force: —

In Fig. 60 we have a view of the northern hemisphere as it is seen from above the north pole. The direction of axial rotation is that shown by the curved arrows along the outer circle representing the equator. This direction is opposite to that of the hands of a watch. Suppose that at A there is a wind blowing along the meridian towards the north pole, as shown by the arrow a. While the meridian at A is advancing to the position B (due to the diurnal rotation of the earth), the air which was at A has been moving in a straight line, that is, in the direction indicated by the arrow a, so that this same direction

[1] The formula expressing the amount of this force is $2\ MVW \sin D$, where M is the mass, V the velocity, W the angular rotation of the earth on its axis, and D the latitude. The meaning of this formula is too complex for explanation in this elementary treatise, but is fully given in Ferrel's "Winds."

when the meridian reaches the position *B* will be indicated by the arrow *b* parallel to the arrow *a*. But it is seen that this direction of the arrow *b* is to the right of the meridian *A* in this new position *B*. Similarly, when the meridian *A* has assumed the positions *C* and *D*, the air will have a motion in the directions shown by the arrows *c* and *d*.

Another set of arrows on the right hand shows the effect of the deflection (due to the earth's rotation) on air motions towards the west; another set of arrows at the top shows the effect on motions from the pole towards the equator; and still another set of arrows on the left shows the effect on motions towards the east.

A revolving terrestrial globe with a fixed brass meridional scale gives a still clearer idea of this matter if one takes a pencil in hand as if for writing, rests the hand, by means of the end of the little finger, on the globe, and then slowly revolves the globe from west to east, letting the hand move with it, but at the same time giving a slight movement to the pencil point (as in writing), and keeps the latter moving in a direction either perpendicular or parallel to the fixed brass meridian scale. The path traced out by the pencil point will show the direction of deviation which the moving air would take for an east-westerly or north-southerly initial direction. Any other directions than the four cardinal ones mentioned can be followed out by making the path of the pencil always keep a fixed angle with the brass meridian scale.

The amount of the influence of the rotation of the earth is best shown by an example. At latitude 50°, a rifle ball moving 1,700 feet per second, discharged at a target 3,300 feet distant, would deviate about 4 inches to the right of the target in the northern hemisphere, and the same amount to the left of the target in the southern hemisphere. This effect may seem slight; but when the force operates on masses of air moving day after day, and over distances of thousands of miles, its results are of great importance.

Effect of the Deflecting Force on the Polar and Equatorial Air Currents. — The qualitative effect of this force on the horizontal meridional motions which have been described, is to cause the free-moving air currents to depart to the right of their course in the northern hemisphere, and to the left in the southern hemisphere.

The upper poleward currents are, then, deflected towards

the east, and the lower currents flowing towards the equator are deflected towards the west, in both hemispheres. We thus see that there will be opposite directions of motions high in the air and near the earth's surface, not only in a north-and-south direction, but also in an east-and-west direction (the upper horizontal current tending towards the northeast, and the lower one towards the southwest, in the northern hemisphere); and the friction, and the intermingling of the air which arises at the boundary of these oppositely directed motions, are very important in their effects. There results then a gyratory motion around the pole, toward it aloft, and away from it below; and, from the law of conservation of areas, the velocities aloft tend to increase, and those below to decrease.

Easterly Motion of the Air in High (North) Latitudes.— Since the effects of this deflecting force accumulate as long as the current continues, therefore in northern latitudes the motion to the right, or easterly motion, of the upper poleward current, becomes very great in comparison with the westerly motion at the beginning of the equatorial current beneath it. So that this upper easterly motion, by friction, and especially by intermingling of the air, gradually entirely overpowers the lower westerly motion, and a general easterly motion results in both the upper and lower layers of the atmosphere in these latitudes. This general eastward air movement, it is found, actually takes place around the pole.

Westerly Motion of Air in Low (North) Latitudes.— In passing towards the equator, however, the lower equatorial current gains strength in its endeavors to flow towards the west, in obedience to the force carrying it towards the right; and after reaching a certain latitude it is able, through friction and the intermingling of air in the currents, to just neutralize the (to that point overpowering)

effects of the upper eastward current, so that there will be no east or west motion in the lower air layers. Passing still farther south, however, the effects of the force towards the west increase, and a westerly air current arises at low altitudes; and as lower latitudes are reached, the power of this western current becomes relatively so great as to overcome the upper easterly current, which is weak there; and in very low latitudes a general westerly motion of the air takes place at all altitudes. So that very near the equator there is a wind directed a little to the north of west in the upper, and a little to the south of west in the lower, air layers.

The Limits of Velocities of Air Currents. — Since the deflecting effect of the earth's rotation is zero at the equator and increases with the latitude, then the poleward upper current is more and more deflected towards the east as it moves northward; and since the cause of the original motion acts along the meridians from the equator towards the north, then the deflection towards the right, due to the earth's rotation, must be directed partly against this movement, and must have a retarding influence on it. This deflecting force increases with the increase in velocity; and it is evident that if the velocities increased enough, there would be a deflecting force towards the right (in this case towards the equator) sufficient to counterbalance the force producing the original poleward current, and this motion would cease. We thus see that there is a practical limit to the velocities which the polar current may attain; for, if these increase too much, the deflection to the right will also increase to such an extent as to actually overcome the poleward current in its course. This acts like a controlling governor on the air circulation.

A second control over excessive air motions is the fact that when a current is impelled by a constantly acting force, the current is continually breaking up into relatively small atmospheric whirls (analogous to the whirls in flowing water) which impede the regular progress of the current by mixing up the various air masses.

A third impediment to excessive velocities is the frequent occurrence of vertical currents in the air.

Vertical Air Movements in the General Circulation. — The vertical currents in the general atmospheric circulation — those which connect the horizontal currents — are the descending current in polar latitudes, and the ascending current in the neighborhood of the equator. Two additional vertical currents shown in Fig. 55 at about latitude 30° are explained on p. 203 in connection with the effects of the air-pressure distribution on the general circulation of the air.

Downward Current at High Latitudes. — The upper air, as it approaches the pole, begins to descend, or settle down towards the earth, and the comparatively great easterly motion which the air has in this region is mostly used up in overcoming the friction between the successive air layers as the air makes its way downward; but aloft there is a sufficiently powerful motion towards the east to give an easterly tendency to the air even at the surface of the earth in this region. Moreover, when the air descends in the higher latitudes, it approaches a little nearer to the axis of rotation of the earth; and its velocity becomes slightly increased, according to the conservation of areas.

Upward Current at Low Latitudes. — The upward current, which extends to a considerable distance from the equator poleward, must in its lower course have a motion towards the west imparted to it by the lower horizontal air current flowing towards the equator. This west component of motion decreases with the altitude up to a certain height above the earth's surface, where it changes into an eastward motion; for at high altitudes there is an easterly current over the whole region from the equator to the pole, except, perhaps, directly over the equator, where the westerly motion extends up to very high altitudes. The altitude where the change of direction takes place de-

creases up to about latitude 30°, beyond which the wind becomes generally easterly (towards the east).

Effect of Air Motions on Barometric Pressure. — The effect of these general motions of the atmosphere on the distribution of the air pressure and the isobaric surfaces is of great importance. If the air temperature were everywhere the same, then the isobaric surfaces would be level. Now, we have seen that by heating the air at the equator the successive isobaric surfaces are elevated there; but directly at the surface of the earth there is no elevation of the isobaric surface, and therefore there is no pressure gradient at the earth's surface such as exists at various altitudes above it. But when there are easterly or westerly air currents at (or close to) the earth's surface, this uniformity of the isobaric surface at the earth's surface no longer exists, but there arise surface gradients; and where there are no east or west motions, there are no gradients at the earth's surface.

The Barometric Gradients at the Earth's Surface are accounted for by the action of the centrifugal force which arises from the eastward (or the westward) movement of the air, this movement being equivalent to a whirling motion around the pole as a center (or, in more local whirls, around some other point as a center). Whenever this whirling motion occurs, the particles so moved tend to leave their curvilinear track, owing to the centrifugal force.

In the case of a whirling mass of water such as is to be seen in any whirlpool, the water particles are free to move, and there is a heaping-up of the water around but away from the center of the whirl, and a decrease of the water towards the center of the whirl. The water recedes from the center to a distance such that the centrifugal force causing it to move away is just equal to

the downward push, or gradient force, of the heaped-up water towards the center of the whirl.

In the case of the air motion around the north pole, which is the center of the whirl, the current towards the east, in the middle and northern latitudes, gives rise to the action of centrifugal force, which causes a heaping-up of the air at the latitudes away from the pole, and consequently to the right and to the equatorward of the easterly current, and a depletion or lessening of the air on the inner, the left or polar, side of the current. Thus we have a lessening pressure towards the pole, just as there is towards the center of a water whirlpool.

The air on the equatorial side is thus thrown off by the centrifugal force, and distributes itself over the equatorial regions of the northern hemisphere, where it meets with a corresponding heaping-up of the air coming from the polar whirl of the southern hemisphere.

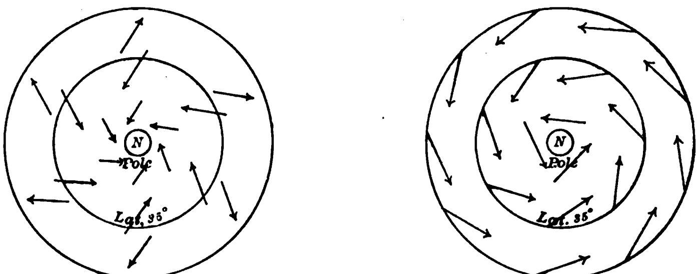
Barometric Pressure Gradients at Various Altitudes. — Up at considerable altitudes above the earth's surface, this heaping-up of the air and increase of pressure take place clear to the equator, as just mentioned; so that at high altitudes there is a maximum pressure at the equator, and a continuous decrease from thence to the pole. But the maximum pressure at the earth's surface is found at about latitude 30°, where the easterly motion ceases, and there is no east or west motion. When, however, the westward motion sets in, as we pass from this point towards the equator, then there is a slight decrease in the pressure, due to the opposing centrifugal effects of the westerly wind. It is not so much a real decrease in the pressure as it is a lessening of the pressure which would exist if the easterly wind alone existed all the way to the equator. This is brought out more plainly in the variation of the pressure

with the latitude in the southern hemisphere (see table, p. 96), where the disturbing influence of the land is not so markedly felt.

The maximum pressure, then, is to be found at about latitude 30° at the earth's surface, but with increasing altitude it lies nearer and nearer to the equator. Up above this dividing line of maximum pressure there is a motion from the west, probably all the way to the equator, and over the lower westerly motion which exists between the equator and about latitude 30°. This allows, then, a motion towards the east up at high altitudes in all latitudes.

Cause of the Equatorial and Poleward Surface Winds in Lower Middle Latitudes. — In the meridional projection of the general air circulation (Fig. 55) the poleward motion is represented above, with the equatorward motion along the surface below it; but there is still a peculiarity which must be mentioned, and which can be best illustrated by the diagram. It is seen that (on the meridional section) at about latitude 30° there is a current along and near the surface of the earth moving towards the pole, which is contrary to what might be expected; and another towards the equator, although this is in the direction to be expected. These currents have as their cause the heaping-up of the air shown by the ring of high pressure at latitude 30°. The atmosphere here exerts a greater downward pressure than that either side of it, and so forces out the air from the lower layers beneath, in the attempt at equalizing the pressure. Part of the air from this lower layer is then forced poleward, and part equatorward. These currents are below the great equatorial return current, and are fed by a downward vertical current in the lower air layers: they are indicated in Fig. 55 at the latitudes marked "Tropical Calm and Dry Belt."

The Air Circulation, viewed as a Whole, can be considered as consisting of two huge atmospheric whirls with the poles as centers (Fig. 61). The direction of rotation of these whirls is determined by the rotation of the earth on its axis; in the northern hemisphere it is opposite to the movement of the hands of a watch (face upwards), and in the southern hemisphere the direction is the same as that of the hands of a watch. At the outer edge of each of these whirls (that is, between the tropical region and the equator) there is an atmospheric ring or belt,



Equator *Equator*

Lower Currents *Upper Currents*

FIG. 61.—GENERAL HEMISPHERICAL DIRECTION OF HORIZONTAL AIR CURRENTS.

decreasing in latitudinal width with increase of altitude above the earth's surface, in which the motion of rotation is in the opposite direction to that of the respective whirls. There is an outward flow of air, at middle altitudes, from the polar centers of these whirls. In the transition zone between this inner (polar) region and outer (equatorial) ring with opposite directions of rotation, there is a heaping-up of the air, and a consequent increase of the air pressure, due to the centrifugal force of these motions. The moving air obeys the law of the conservation of areas (see p. 182).

An Interchange of Air between the Northern and Southern Hemispheres also takes place, owing to the transference of the region of greatest temperature, first to one side, and then to the other side, of the equator, with the annual movement of the sun. When, for instance, we have our winter in the northern hemisphere, and summer exists in the southern hemisphere, then the isobaric surfaces are more elevated by the intenser heat and subsequent expansion of the air over the equatorial regions in the southern hemisphere than in the northern hemisphere, and air flows down these surfaces from the southern into the northern hemisphere. Similarly, when we have our summer in the northern hemisphere, there is a flow of air across the equator from the northern into the southern hemisphere.

Inequality in Air Motions Due to Variations in Temperature. — The difference in temperature between the polar and equatorial regions is much greater in the winter time than in the summer time, and consequently the slopes of the isobaric surfaces are greater; and since the velocities of the air currents depend on the gradient or slope of these surfaces, these velocities are much greater in winter than in summer. From the observed data of wind velocities it is seen that the winter velocities are in many cases double those of the summer.

Theoretical Computation of Easterly and Westerly Velocities of the Wind. —Attempts have been made to compute theoretically the easterly and westerly velocities of the wind at various altitudes on the different parallels of latitude, by means of the gradient forces resulting from the inclination of the isobaric surfaces due to differences in temperature and air pressure between the equator and the poles along an average meridian. These velocities in miles per hour, for the yearly average, are given on p. 206.

It is seen from this table that the westerly motions disappear at a height of about 10,000 feet above sea level to the poleward of about

COMPUTED EASTERLY OR WESTERLY WIND VELOCITIES ALONG A MERIDIAN.

Latitude.		Easterly (E.) or Westerly (W.) Velocity of Wind in Miles per Hour, at Various Altitudes.			Increase in Easterly Velocities with each (about) 3,300 Feet in Altitude.
		Sea level.	About 3,300 feet.	About 13,200 feet.	Miles per hour.
N. Lat.	75°	W. 2.7	E. 0.2	E. 9.2	E. + 3.0
	70°	W. 2.0	E. 2.0	E. 14.3	E. 4.1
	65°	E. 0.1	E. 4.9	E. 19.3	E. 4.8
	60°	E. 2.4	E. 7.6	E. 23.1	E. 5.2
	55°	E. 3.4	E. 8.7	E. 24.5	E. 5.3
	50°	E. 3.3	E. 8.7	E. 24.9	E. 5.4
	45°	E. 3.0	E. 8.5	E. 25.0	E. 5.5
	40°	E. 1.6	E. 7.2	E. 24.0	E. 5.6
	35°	W. 0.7	E. 5.0	E. 22.4	E. 5.8
	30°	W. 5.3	E. 0.6	E. 18.2	E. 5.9
	25°	W. 8.9	W. 3.1	E. 14.4	E. 5.8
	20°	W. 9.4	W. 3.8	E. 13.0	E. 5.6
N. Lat.	15°	W. 7.8	W. 4.3	E. 6.1	E. 3.5
Equator	0°				
S. Lat.	15°	W. 15.6	W. 10.5	E. 4.8	E. 5.1
	20°	W. 13.0	W. 8.2	E. 6.4	E. 4.8
	25°	W. 6.4	W. 1.7	E. 12.5	E. 4.7
	30°	E. 2.4	E. 7.0	E. 21.0	E. 4.7
	35°	E. 7.7	E. 12.3	E. 26.1	E. 4.6
	40°	E. 11.6	E. 16.2	E. 30.0	E. 4.6
	45°	E. 14.9	E. 19.5	E. 33.3	E. 4.6
	50°	E. 17.1	E. 21.7	E. 35.7	E. 4.6
	55°	E. 17.0	E. 21.6	E. 35.6	E. 4.7
S. Lat.	60°	E. 13.6	E. 18.2	E. 32.2	E. + 4.7

latitude 15°. And we see here, derived by theory, the observed fact of the easterly wind velocity near the earth's surface in the northern hemisphere reaching a maximum at about latitude 50° or 60°; but it is to be noticed that this region of greatest velocity decreases in latitude with increase of altitude.

Observed Circulation of the Atmosphere.—Usually the arrows showing the direction of the winds on charts are drawn partly from direct observation, and partly from the known laws of the direction of wind as depending on the distribution of the air pressure. Such are the indications of wind directions given on the charts of air pressure for the globe (see Figs. 26, 27).

The lower air currents are difficult to observe on the continents, owing to the irregularities of the ground, and so we must turn to the observations made on the ocean for the best obtainable presentation of their characteristics. Published charts showing the general wind direction for the Atlantic Ocean have been already given (see Figs. 36, 37). It is seen on these, that while the surface wind directions follow in general those deduced by theoretical reasoning and shown in the previous sections, yet there are localities in which the local pressure conditions exert such a predominating influence that the main currents are interrupted. Thus, on the northern North Atlantic and middle South Atlantic oceans, there are local atmospheric whirls formed in the winter of the northern hemisphere; and in the summer there is still another interrupting whirl formed on the middle North Atlantic.

In the northern hemisphere, the observed general direction of the upper air currents is from N. 45° W. to S. 45° E., in high latitudes like Lapland. In middle latitudes (at about latitude 40°) the direction is nearly due easterly. In lower latitudes (at about latitude 25°) it is perhaps from S. 75° W. to N. 75° E. In the equatorial regions, however, the direction is in general from the east towards the west, and in some cases it is from S. 45° E. to N. 45° W.; but there is a much greater irregularity of direction in this region than near the poles, on account of the shifting of the region of greatest heat from one hemisphere to another. There exist comparatively few wind observations, on the land, for the intertropical regions.

Trade Winds. — These are the lower winds belonging to that part of the general atmospheric circulation embraced between the equatorial region of calms and the tropical belt of high atmospheric pressure. They blow from the northeast in the northern hemisphere, and from the southeast in the southern hemisphere. The general location and direction of these winds are shown on the diagram of the general circulation of the atmosphere (see Fig. 55), where they are indicated by heavy arrows between the Tropical and Equatorial Calm Belts. They have not very great velocities, but they are relatively very constant in their continuance and in their velocity of motion. The whole system of trade winds shifts from north to south, and the reverse, following the sun in its course; and it also varies somewhat in width, not only with the longitude, but also with the season of the year, the latter being shown by the following table: —

LATITUDINAL LIMITS OF THE TRADE WINDS.

	AVERAGE OVER THE PACIFIC AND ATLANTIC OCEANS.	
	IN MARCH.	IN SEPTEMBER.
In northern hemisphere (northeast trades) . .	Lat. 4° N. to 25° N.	Lat. 10° N. to 33° N.
In southern hemisphere (southeast trades) . .	Lat. 1° N. to 26° S.	Lat. 5° N. to 22° S.

Doldrums. — Between the surface winds of the two hemispheres there is an equatorial region of calms, called the *doldrums*. This region lies a little to the north of the terrestrial equator; but it varies somewhat in locality and width, not only regularly with the season of the year, but also irregularly with the longitude. In March the dol-

drums extend from 0° to 3° north latitude on the Atlantic Ocean, and from 3° to 5° north latitude on the Pacific Ocean; in September they extend from 3° to 11° north latitude on the Atlantic Ocean, and from 7° to 10° north latitude on the Pacific Ocean.

Monsoons, or continental land and sea winds, occur during the hot and cold seasons of the year. During the summer the interiors of the continents become heated, and this elevates locally the surfaces of equal air pressure, and the air flows down these towards the ocean at some distance above the ground; and a countercurrent near the ground sets in from the ocean towards the land. In the winter time the opposite interchange of air takes place, and along the ground the air flows from the land towards the sea.

In case of an uninterrupted low-level region extending from the ocean inland, the monsoon effects are not so marked as when the interior is high or mountainous. One reason for this is, that where there is a long surface slope the conditions are best for an extended general movement of the air. In winter the lower air will flow downward much easier than on a level; and in summer the warm air flows upward along the slopes more easily than along a level surface, just as it will rise better in a somewhat vertical flue than move in a horizontal flue; and there is less tendency for local currents to form. Another reason is the fact that the temperatures on elevated land, especially on plateaus, are greater in summer and less in winter than at the same altitudes above the ocean or the low land; and there is thus a contrast in the temperatures maintained to a height depending on the altitudes attained by the land. This principle is illustrated by the greater air circulation (or draught) in the case of a tall chimney as compared with a short one. This effect is felt mostly in the summer monsoon, as in winter the air temperature of the land is not so much below that of the surrounding air as it is above in summer.

While the monsoon winds occur in many regions, yet nowhere else are the conditions so favorable as in India and the north Indian Ocean. There the low lands are backed by the Himalaya mountain range, to the north of which lies the Kuenlun range, with the plateau of Tibet

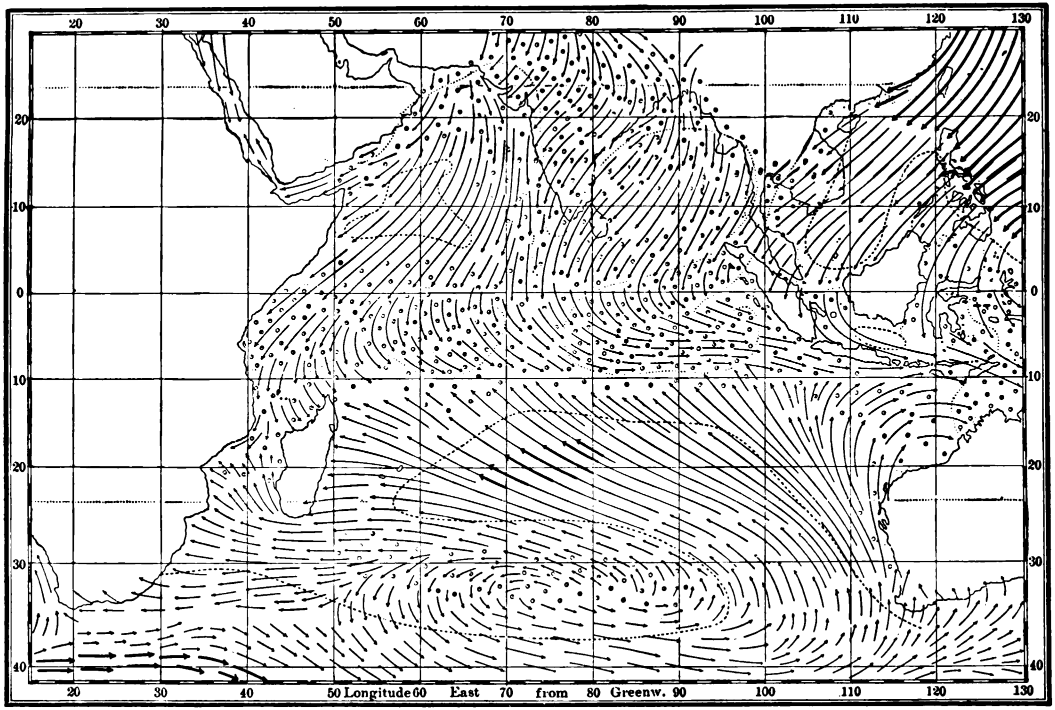


FIG. 62.—WINDS OF THE INDIAN OCEAN, JANUARY AND FEBRUARY (DEUTSCHE SEEWARTE).

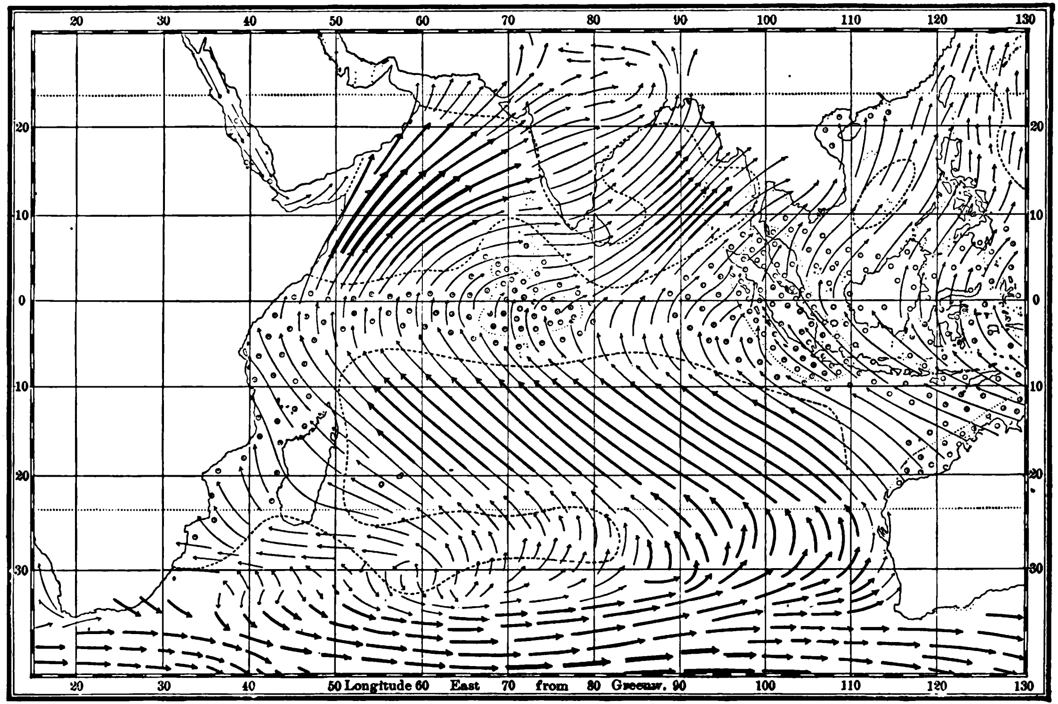


FIG. 63.—WINDS OF THE INDIAN OCEAN, JULY AND AUGUST (DEUTSCHE SEEWARTE).

between; and still farther to the north lie vast stretches of quite high plateau desert land. During the summer monsoon the region of great surface heat of the Indian Ocean and the adjacent low lands, which consequently affords the air an excessive capacity for moisture, lies close to the elevated region just described; and when the air moves up the steep slopes, as a result of the adiabatic cooling excessive precipitation takes place, and enormous quantities of freed latent heat increase the temperature of the air over the continent above what it would be in the case of no precipitation, and the wind velocity is thereby increased.

Strongly developed monsoon winds so act that they greatly influence, and may even reverse, the direction of the regular primary wind circulation.

The accompanying figures (Figs. 62, 63) show the very powerful monsoon winds of the north Indian Ocean and southern Asia in midwinter and midsummer. Heavy arrows indicate strong winds; light arrows, light winds; and circles, calms. Regions of variable winds are denoted by short arrows, while the long arrows designate steady winds. In January and February the outflow of air from the interior of the Asiatic continent is southward across the north Indian Ocean to 10° of latitude beyond the equator. In July and August there is an inflow of air towards the interior of the Asiatic continent, not only extending to the equatorial region of the Indian Ocean, but also forming a continuation of the winds of the lower latitudes of the southern hemisphere (to 30° south latitude) which flow in towards the equator. This continuous northward sweep of the winds over the ocean from 30° south latitude to over 20° north latitude, caused by the great indraught over the heated continent, readily accounts for the excessive force of the southern monsoon winds of midsummer on the Indian seas and mainland.

CHAPTER IX.

SECONDARY CIRCULATION OF THE ATMOSPHERE.

We must next examine those more local conditions of the atmospheric circulation which interrupt the general circulation, and occasion the variability in the winds during brief periods of time.

We will take up those conditions which are most closely connected with, and to a great extent depend on, the currents of the general air circulation. Such are the local atmospheric whirls which occur irregularly in the greater hemispherical currents, and which appear to gradually form in the greater current, follow it in its general course for a time, develop in intensity to maturity, and then gradually disappear.

Movement of Air in Whirls. — It is a marked characteristic of atmospheric motion that it occurs mainly in the form of whirls or vortex motion. Even when the wind blows in an apparently straight line, if it is followed out far enough, it will usually be found to belong to some system of whirling motions. Thus the great air currents forming the general circulation of the atmosphere have been found to be but a part of the great hemispherical vortical movement described in the preceding chapter.

Secondary Atmospheric Whirls. — Within these mighty air currents of the hemispherical air circulation there exist limited systems of atmospheric whirls in which, in some cases, the air moves spirally inward towards a center, and

in others spirally outward from a center. In such local whirls there exist gradients of air pressure such as occur in the great hemispherical whirl: and these may be arranged in two ways, — first, there may be an area of lower atmospheric pressure at the center of the whirl, in which case the air movement is spirally inward, and such a system is called a *cyclone;* second, there may be a region of higher atmospheric pressure at the center, in which case the spiral air movement is outward from the central area, and such a system is called an *anticyclone* (Fig. 64).

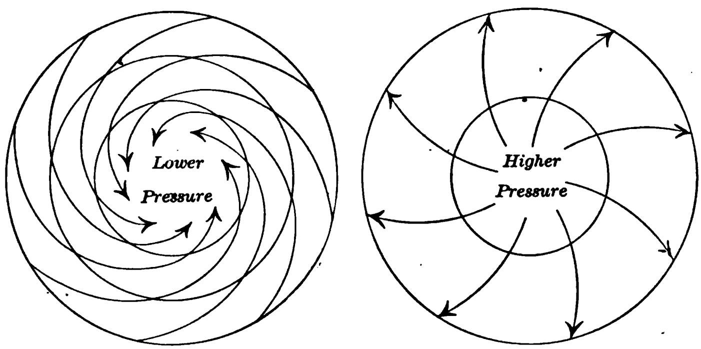


FIG. 64. — DIRECTION OF WHIRL FOR A CYCLONE AND AN ANTICYCLONE (NORTHERN HEMISPHERE).

Near the center of a cyclone there is an upward movement of the air, while in an anticyclone there is a central downward movement.

Cyclonic conditions may be characterized thus: a central region of low air pressure, with a gradual increase from thence toward the outer limits; a spiral motion of the lower air inward toward the center; and in our middle latitudes, on the eastern and southern sides of the center of cyclones, a warm, cloudy, more or less rainy condition, with strong winds from the east and south, and on the

western and northern sides a cool, clear, or clearing condition, with strong winds from the west and north.

Anticyclonic conditions may be characterized thus: a central region of high air pressure, with a gradual decrease from thence to the outer limits; a spiral motion of the lower winds outward from the center; and in our middle latitudes, on the eastern and southern sides of the region of highest air pressure, a cool, clear, or clearing condition, with winds from the north and west, and on the western side a warm condition, with increasing cloud, and winds from the south and east.

These whirls have in our latitudes a motion of translation of about 30 miles per hour, usually, in an easterly direction; and as they pass over any place lying in their path, first the conditions on the eastern or front side are experienced, and then those on the western or rear side.

A cyclone or an anticyclone may sometimes assume such large proportions as to cover half the United States, and may exist for several days, and cross the entire country.

The direction of the inward spiral movement of the air circulating around an extended cyclone is, in the northern hemisphere, always opposite to that of the hands of a watch; and in the southern hemisphere it is with the hands of a watch. The direction of the outward spiral movement of the air circulating around an anticyclone is always with that of the hands of a watch in the northern hemisphere, and opposite in the southern hemisphere.

Whether the air is flowing down the isobaric surfaces inward towards the center of low barometric pressure of the cyclone, or outward away from the center of high barometric pressure of an anticyclone, the rotation of the earth on its axis causes the current of air to depart to

the right in the northern hemisphere, and to the left in the southern hemisphere. This gives the distinctive directions to the whirl of the air masses in cyclones and anticyclones.

These cyclones and anticyclones are sometimes of so great extent, and they are so constantly undergoing changes, that, in order to study such in their true relations, it is necessary to obtain simultaneous observations of the meteorological elements over extensive regions, such as continents, oceans, or even a whole hemisphere.

CYCLONES.

Classes of Cyclones. — The term *cyclone* is applied by meteorologists to all kinds of atmospheric disturbances in which the air pressure decreases, and there is a wind movement, inward towards the center. They are not, however, all alike in their other characteristics.

There are the cyclones of the torrid zone, which in their greatest development are called *hurricanes* and *typhoons*. These atmospheric whirls vary from a few miles to several hundred miles in diameter, and are characterized by great violence of wind and copious rainfall.

Cyclones of the temperate and colder zones are of greater extent, usually covering a region at least several hundred miles, but sometimes a couple of thousand miles, in diameter. These are of less intensity than the cyclones of the torrid zone, and are the *cyclonic areas*, or areas of low barometer, usually spoken of in our government weather reports.

The latter cyclones are frequently accompanied by secondary cyclones of limited extent but intense energy, manifested by strong winds. These are called *tornadoes*.

The cyclones of the torrid zone sometimes pass north into the temperate zone, where they gradually lose their former characteristics of relatively limited extent and great energy, and spread out and assume the main characteristics of the cyclones of the temperate zone.

Characteristics of Cyclones. — We do not as yet understand to a certainty the process of formation of cyclones, and so shall at first describe their characteristics, and afterward mention their probable causes.

The main features of the cyclonic areas which must be studied are the distribution of air pressure, air temperature, moisture, cloud, and rainfall; the direction and velocities of air currents or winds; and finally the direction and velocity of propagation of the cyclonic areas, and the paths usually pursued by them.

Distribution of Air Pressure in Cyclones. — In cyclones, at the earth's surface and in the lower layers of the atmosphere there is a central area of low air pressure from which there is an increase of the pressure in all directions (see within the dotted circle, Fig. 65). This is best represented by means of isobaric lines drawn for equal intervals of barometric pressure (after the observed air pressures have been reduced to sea level). Such isobaric lines show where the isobaric surfaces intersect this level. The isobars in the simplest ideal case are concentric circles, the inner one of which has the lowest pressure, the outer ones having successively increasing pressures; but the isobaric lines seldom present these regular features, and in nearly every case they are elongated.

The ratio of the longer axis to the shorter may be more than 4 to 1; on the continent of America it averages about 2 to 1; while on the Atlantic Ocean the average is only 1.7 to 1.

The longest axis in our middle latitudes usually lies in a direction from southwest to northeast, and its average direction for a great many cyclones in the United States is N. 36° E. The barometric gradient, or rate of change

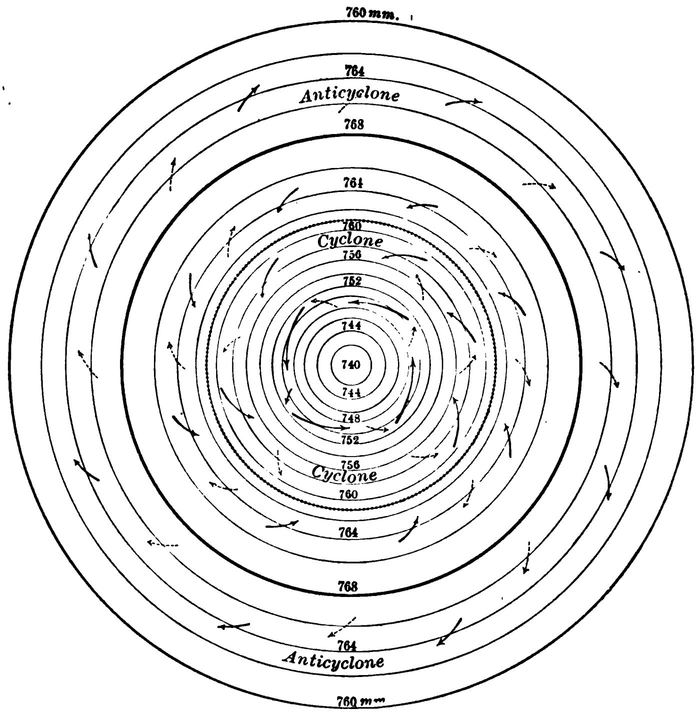


FIG. 65.—AIR PRESSURE AND AIR MOTION WITHIN A CYCLONIC AREA (AFTER FERREL).

of air pressure, is steeper or more rapid, and the isobars lie closer together, over the ocean than over the land.

Where the isobars are considerably elongated, — that is, are very elliptical, — there are sometimes two or more distinct centers of low pressure within the cyclone; but they

differ only slightly from each other in pressure, and frequently have the same pressure.

The increase of air pressure from the center of cyclones to the extreme outer regions is not a wholly uninterrupted ascent; for at some distance from the center there exists a ring of higher air pressure, with anticyclonic characteristics (see Fig. 65), which is especially noticeable in cyclones of great energy, such as the hurricanes of the torrid zone; and in cyclones of great extent it lies in the region of regular anticyclones. This ring of high pressure is similar to that at about latitude 30° in the hemispherical whirl. The barometric pressure at the center of the cyclone sometimes descends as low as 27.5 inches.

Fig. 68, *a*, on p. 222, shows in an ideal cyclone the isobars at sea level and at an altitude of several miles.

Distribution of Temperature in Cyclones.—The rear and left-hand sides, facing the direction of translation (western and northern sides of our cyclones), are the coldest; and the front and right-hand (southern and eastern) sides are the warmest. But where the winds blow from over a water surface, these conditions may be modified considerably, since the air would then be warmer in winter and cooler in summer than that from over a land surface. In the central and eastern United States there is sometimes a difference of over 70° F. in the temperatures on the warmer and colder sides of cyclones; and the isotherms which in their normal condition run east and west will become so changed as to run north and south.

The following diagram (Fig. 66) shows the isotherms in a cyclone of, say, 500 miles in diameter, as indicated by the circle, along which also the direction of the wind is shown by means of arrows.

With the increase in the altitudes there is (at least over the interior of continents) a rapid decrease in the temperatures, because in the lower air layers the temperature is relatively high, while above it is relatively low. We need still more observations on this point, however, before definite laws can be formulated.

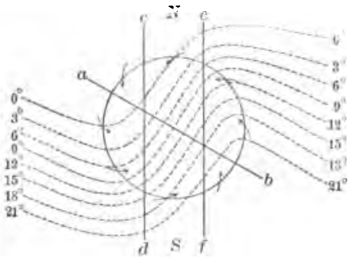
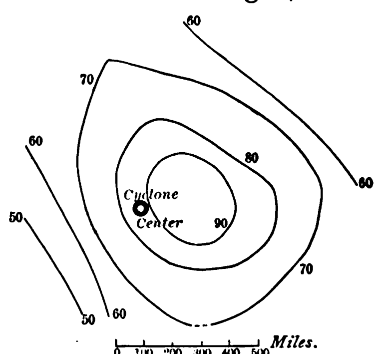


FIG. 66. — ISOTHERMS IN A CYCLONE. C°. (AFTER FERREL.)

Distribution of Moisture, Cloudiness, and Rainfall in Cyclones.— These elements are so intimately connected that they may be considered together. The quantity of moisture in a cyclone depends so much on its location with respect to the main source of moisture, the sea, that no general laws can be enunciated concerning it, or the cloudiness and rainfall resulting from it. In our latitudes the eastern, southeastern, and southern sides of the cyclone have the greatest capacity for moisture, because they are the warmest; and generally the western and northern sides are the freest from cloud. The various kinds of cloud have such different origins, and lie at such different altitudes above the earth's surface, that their individual distribution in the cyclone is difficult to give with accuracy. The cirrus



O Signifies the center of the cyclone. The other lines show the degree of cloudiness in percentage of an overcast sky.

FIG. 67. — CLOUD DISTRIBUTION IN CYCLONES (AFTER CLAYTON).

clouds, however, are usually to be found on the front side of a cyclone, and extend considerably in advance of it.

The relative humidity increases towards the center of the cyclone, being greatest on the eastern and southeastern sides.

The distribution of cloud around the center of a cyclone in the eastern United States is shown in the preceding diagram (Fig. 67), where the top is the northern side.

The Rainfall occurs most frequently on the side of the cyclone which is nearest the supply of moisture, whence this last is carried into the region of the cyclone by the winds blowing in towards the center of the cyclone.

The great downpours of rain, however, occur at the front (eastern) or warmer side of our cyclones two or three times as frequently as in the cooler (western) side.

The cyclones which are accompanied by excessive rains have strongly marked characteristics, and have very low air pressures at the center; while those which have light rainfall, or even none at all, have feebler cyclonic characteristics, and but slightly depressed air pressures at the center.

The areas within which rain falls at any one time are usually elliptical in shape, and in general are twice as long as they are broad; and sometimes they are as much as 2,000 miles in length. It is usually found that there is a central area of maximum rainfall, and this is encircled by successive zones of diminishing amount. In the eastern and northern parts of the United States the area of maximum rainfall lies southeast of the center of the cyclone, and usually at a distance of about 300 miles from it; but the distance varies greatly in individual instances.

Cyclones are of more or less gradual development.

The greatest depth of barometric pressure does not occur at first, but is reached later; and the maximum amount of rainfall occurs at the time of minimum pressure.

Direction and Velocities of Winds in Cyclones. — As the air moves spirally inward towards the center of low barometric pressure, it intersects the successive isobaric lines at certain angles, which vary not only on the different sides of the cyclone and at different distances from the center, but also over a land and over a water surface. Fig. 65 shows this circulation at the earth's surface, and

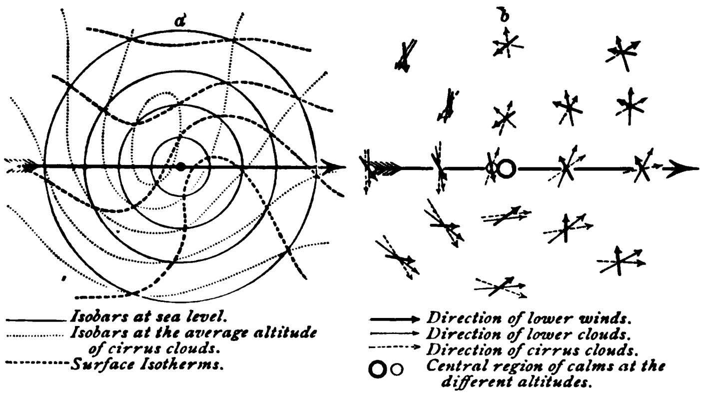


FIG. 68. — ISOBARS, ISOTHERMS, AND WINDS AT VARIOUS ALTITUDES IN A CYCLONE. (Köppen.)

aloft, from the outer to the inner limits of the cyclone. The wind cuts the isobars at a smaller and smaller angle as the center is approached; that is, it turns away more and more from the center as it moves inward, and finally close to the center flows nearly along the isobars.

The direction of the wind circulation around a cyclone at various altitudes is shown in the accompanying diagram (Fig. 68, *b*), in which the arrows fly with the wind. The heavy unbroken arrows show the

direction of air motion near the ground; the light unbroken arrows, that at the height of the ordinary lower clouds, at perhaps an altitude of from 6,000 to 9,000 feet; and the broken arrows, that at the height of the cirrus clouds, at an altitude of perhaps from 20,000 to 30,000 feet.

The arrows in Fig. 68, *b*, show that in the lower air layers the direction of the wind is towards the interior of the cyclone, at a moderate altitude it is tangent to the isobars, and at a high altitude it is away from the cyclone. This relation expressed more concisely is as follows: the direction of the air movement around a cyclone in the northern hemisphere departs more and more to the right with increase of altitude.

The wind velocity increases from the outer edge towards the center of a cyclone, but the maximum velocity is reached at some distance from the center (several hundred miles in some cases); and then the velocities decrease again, sometimes with great suddenness. Directly at the center there is usually a calm, or but very little horizontal motion of the air. The velocities at the place of maximum wind depend on the intensity of the cyclone and the depth and steepness of the gradients, and even near the ground frequently reach 40, 50, or 60 miles an hour.

Magnitude of Cyclones. — The usual diameter of cyclones over the United States is from 1,000 to 1,500 miles, while over the North Atlantic Ocean it is a few hundred miles more. The barometric gradients are about 15% steeper over the North Atlantic than over the United States, and this is perhaps due to the greater friction of the lower air layers over the land than over a water surface. The cyclones thus develop more force over the ocean than over the continents.

Direction and Velocity of Movement of Cyclones. — Cyclones, in general, move in the direction of motion of the

FIG. 69. — PATHS OF NUMEROUS CYCLONES IN THE NORTHERN HEMISPHERE (AFTER LOOMIS).

great mass of air carried by the primary atmospheric currents.

The preceding chart (Fig. 69) shows the paths pursued by a great many individual cyclones. The arrow-headed curved lines show at their beginning the place of formation, their length shows the path pursued, and the points of the arrows show the direction of motion and place of dissipation of individual cyclones.

The direction of translation of cyclones for the middle latitudes is easterly, while for the lower latitudes it is westerly, corresponding to the direction of the air currents in the general atmospheric circulation at those latitudes.

The cyclones of our middle latitudes have an average direction of translation of about N. 80° E., or 10° north of an easterly direction; but this varies somewhat for different regions. In the central United States the direction is a little south of east, and in the eastern part it is a little north of east.

The cyclones which are confined to our tropics have a direction of about N. 64° W.; and the Asiatic or East Indian cyclones have a velocity of motion of about 9 miles per hour, while those of America or the West Indies move about 12 miles per hour.

The cyclones which first appear in the tropics with a westerly motion, and then cross over into middle latitudes and move in an easterly direction, have in the West Indies at first an average direction of N. 64° W. and a velocity of 17 miles per hour, and in middle latitudes a direction of N. 52° E. with a velocity of 20 or 21 miles per hour; while those in the East Indies have at first an average direction N. 52° W. and a velocity of 8 miles per hour, and later in the middle latitudes a direction of N. 55° E. and a velocity of 10 miles per hour.

There is a tendency for these cyclones to pursue somewhat the same tracks, according to the place of origination

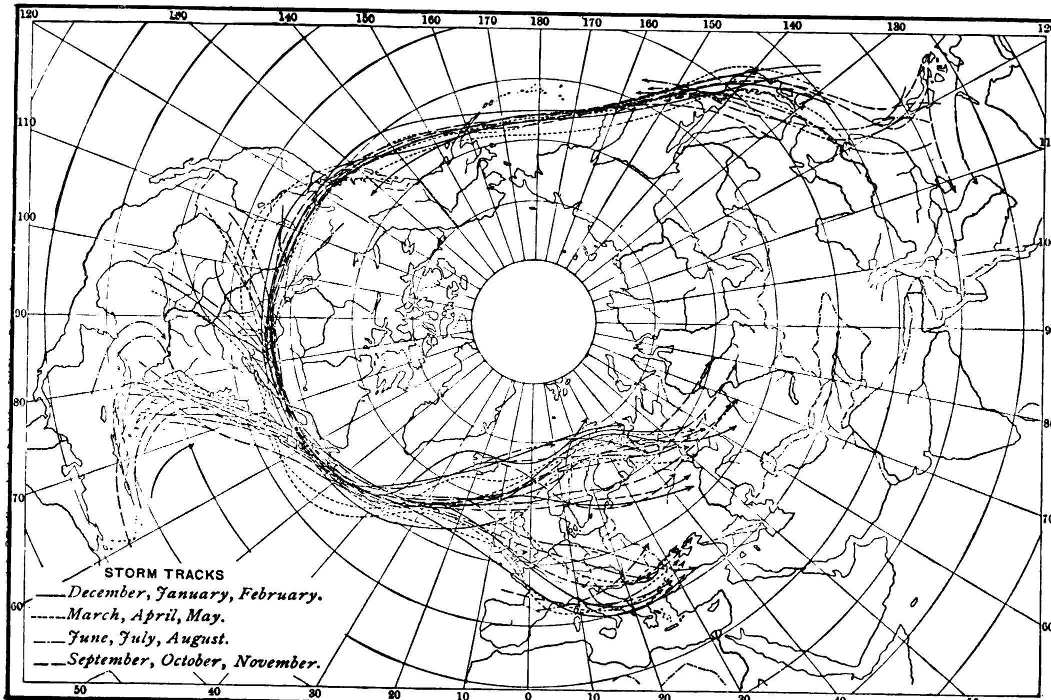


FIG. 70.—AVERAGE TRACKS OF CYCLONIC AREAS, BY MONTHS AND SEASONS, OVER THE NORTHERN HEMISPHERE, 1878-87 (U.S. WEATHER BUREAU).

and the special characteristics of the individual cyclones. Sometimes two cyclones coalesce and form a single cyclone, and at other times a single one will divide up into two, and these will pursue quite distinct paths.

The average tracks of observed cyclones over portions of the northern hemisphere are shown in Fig. 70.

The velocity of translation of cyclones is nearly twice as great in winter as in summer, and seems to increase up to a middle latitude, and then decrease again. The cyclones of the United States move with an average velocity of about 30 miles per hour, which is considerably faster than that of cyclones of the Atlantic Ocean and Europe.

The Seasons of Greatest Frequency of Cyclones. — The cyclones of the West Indies and the China Sea occur most frequently in July, August, September, and October; those of the Java Sea and the South Indian Ocean, in December, January, February, March, and April (in the latter months for the regions more distant from the equator); those of the Arabian Sea and Bay of Bengal, from April to June, and again in October and November. In the middle northern latitudes the cyclones occur most frequently probably in February, and least frequently in July.

Region of Maximum Number of Cyclones. — The region over which the maximum number of cyclone tracks have been traced in the northern hemisphere lies along the following course: commencing in the region south of Japan and Corea, it passes thence through Bering Sea to the southern part of Alaska; thence along the coast almost to Oregon; thence nearly due east across the continent to the southern part of Newfoundland; thence east-northeast over the Atlantic Ocean to the Orkney Islands, and

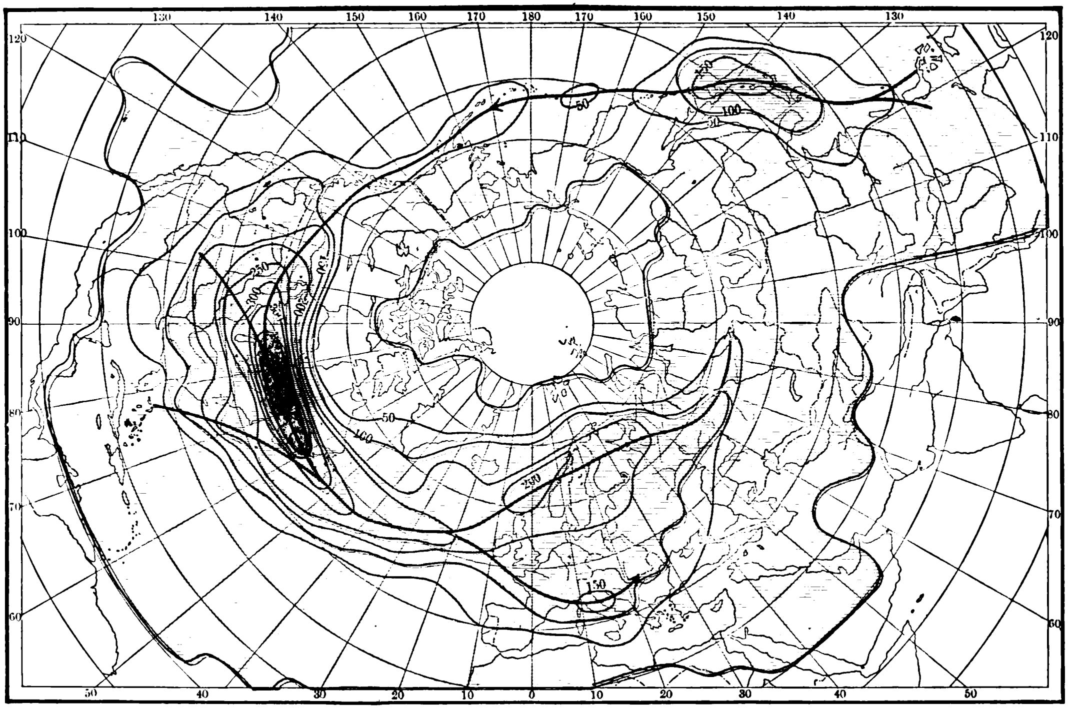


FIG. 71. — CYCLONE FREQUENCY OVER THE NORTHERN HEMISPHERE, 1878–87 (U.S. WEATHER BUREAU).

farther on to northern Norway and Sweden; thence southeast to the interior of Russia and Siberia, in which dry region they probably die out for lack of the moisture in the air which seems necessary for their long continuation.

The freedom of the region within the Arctic Circle from cyclones is due partly to the low temperature and smaller amount of moisture there, and partly also to the descending air currents of that region, and to the absence of the great horizontal currents of middle latitudes.

The frequency of occurrence and region of maximum number of cyclones in various parts of the northern hemisphere are shown in Fig. 71, where the numerals signify the number of cyclone centers which have passed, during a period of 10 years, over the regions along the lines. The heavy arrow lines show the most frequented paths.

Hurricanes and Typhoons. — The hurricanes of the West Indies and the typhoons of the East Indies are cyclones possessing both great extent and power. Commencing usually between latitudes 10° and 20° N., they pursue a course a little to the north of west until they reach latitude 30°, when, if they last long enough, they swerve around towards the east, and pursue a course north of east. They diminish in intensity with their northward progress, and finally enter upon the path frequented by the cyclones of middle latitudes, and disappear in the same manner as these.

The time which it takes for a hurricane to pass over a given place lying within its path varies from a few hours to a day or more, when we count the time during which the barometer is falling as the hurricane approaches, and rising as it recedes from the place. Hurricanes are never more than a few hundred miles in diameter; but they have long paths, and thus their force is distributed over a large territory.

As the hurricane is approaching, there is a decrease of pressure, a slight decrease of temperature (during the

western or northwestern progress), a rapid increase of wind velocity, and a quite constant wind direction. At the center there is almost a calm. At the rear of the center there is an increase of pressure, an increased temperature, first an increase and then a decrease of the wind velocity; and the wind blows quite constantly from a direction nearly opposite to that on the front side. The barometric pressure sometimes falls two inches at the center of one of these cyclones.

One typhoon was traced from near Manilla, where, on Sept. 27, 1882, it had a movement forwards of but 5 miles per hour, to the coast of Japan (about Oct. 1), where it moved at the rate of 33 miles per hour, which increased to 51 miles per hour on Oct. 3, just east of Japan; thence it passed to the Aleutian Islands and to Oregon, which it reached on Oct. 10; thence across the Rocky Mountains at a rate of 37 miles per hour, and through the northern United States and Canada, Hudson Bay, and Labrador, to Davis Strait; and thence past the southern point of Greenland to longitude 27° west, latitude 55° north, in the Atlantic Ocean, where it united with another cyclone. After this union, the cyclone remained stationary for nearly a week (from Oct. 19 to Oct. 24), when it suddenly took a southeast course towards England, passed over the Bay of Biscay, and reached France on Oct. 27, took another northeasterly trend, and vanished in the region of the Baltic Sea on Nov. 1, having thus traveled over 14,000 geographical miles in 35 days.

Eye of the Storm. — Directly at the center of the cyclone there is a region in which the air may be quite free from cloudiness, and in which there is but little horizontal air motion. This is undoubtedly due to the existence of a descending current at higher altitudes above the ascending current at the interior of the cyclone at low altitudes. The descending current prevents the ascending cu[illegible]ching a sufficient altitude to cause pre[illegible]haps even the formation of clouds,

and tends to dissipate clouds already formed. This phenomenon becomes so marked, in the case of the limited but intense tropical cyclones, that the limits of the clear space at the center can be distinctly seen, and it is called the "eye of the storm." In the cyclones of higher latitudes it is not so pronounced in character, but is of wider extent.

Secondary Cyclones sometimes occur within the boundaries of our larger cyclones, usually to the southeast of the center of the main cyclone, where the air is warmer and moister than in other quarters. They have an air circulation against the hands of a watch, like the primary cyclones. Sometimes they are so slight as to cause merely a small distortion of the isobars of the main cyclone, with but little disturbance of the wind direction; and sometimes they cause a fairly well developed cyclonic wind circulation. The local upward currents of these secondary cyclones cause an increase in the cloudiness, and more rainfall at the points where they exist. Tornadoes and kindred phenomena are closely allied to the secondary cyclones, and it is scarcely possible to draw a dividing line between them. It has been deemed best, however, to treat of tornadoes separately in the next chapter.

Origin of Cyclones. — When air flows along in a current, there is a tendency for it to break up into whirls or eddies on the outer edges of the current, and so, near the borders of the great currents of the primary air circulation, such vortices are formed (see Fig. 72); and the centrifugal force resulting from this local whirling motion causes the air to recede from the center of each whirl, and to be heaped up on the outer edge, making a more or less regular increase of air pressure (gradient) from the center to the outer edge of the whirl. These whirls might form the cyclones, or areas of low air pressure; and the inter-

vening spaces, where the air is heaped up, the anti-cyclones, or areas of high air pressure. This theory, however, has not been worked out with any degree of completeness.

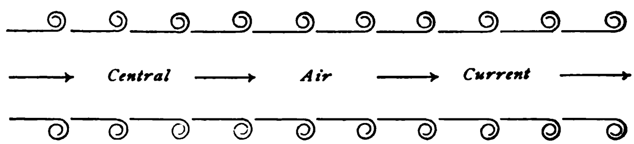


FIG. 72. — BREAKING-UP OF AN AIR CURRENT INTO WHIRLS.

Another Source of Cyclonic Motion is to be found in the local temperature conditions. If anywhere, as in the interior of the continent, a local region becomes excessively heated, the air expands upward, and the isobaric surfaces become elevated, and cause an outflow of air aloft into the colder surrounding air. The increased air pressure in this outer region causes the air near the surface of the ground to flow inward toward the warm central area, where the air pressure has been diminished by the overflow of air aloft. Connecting these two horizontal currents, there is an ascending air current over the central warm region, and a descending current on the outer cooler boundary.

The direction of these horizontal currents, where they extend over hundreds of miles, is affected by the deflecting force of the earth's rotation. The upper outflowing current and the lower inflowing current are deflected towards the right (in the northern hemisphere); and so, instead of having a flow of air directly from and toward the heated center, there results a curvilinear or spiral motion in both currents.

The inflowing surface current is deflected towards the right of the central area, and flows spirally around it in the direction opposite to

that of the hands of a watch (face upwards). As this air current approaches the center, the velocities are much increased, because the same amount of air has to pass in a given time through a much narrower space nearer the center than farther away from it. The horizontal motion becomes more and more changed to the vertical as the immediate center is approached, and near the center becomes an upward current; and this upward current, too, may have great velocities.

The upper horizontal current flowing away from the center is likewise deflected towards the right, and its spiral motion is in the same direction as that of the hands of a watch (in the northern hemisphere); and since the channel for the air constantly increases with the distance from the center, the velocities of the current decrease, and on the outskirts of the whirl there is but a gradual settling-down of the air, instead of a swift down-moving current.

The lower inflowing current is much more clearly marked than the upper outflowing current, and its gyratory motion is so great that the opposite gyratory motion of the upper current is at first required to overcome this lower gyratory motion before it can assert itself in its proper direction; hence, in the neighborhood of the warm center, the gyration of the whole air mass is against the hands of a watch.

As the upper outward current increases its distance from the center, the effect of the deflecting force becomes greater, and finally on the outskirts of the whirl overcomes the less developed deflecting forces of the lower current, and imparts to the whole air mass its distinctive direction of rotation.

Therefore, for a cyclonal region of this kind, nearly the whole mass of air at the interior has a circulation around the center in a direction opposite to that of the hands of a watch; while outside of this there is a ring or zone in which most of the air has a distinctive rotation around the same center, but in the opposite direction to that at the interior. Since the friction with the ground is such as to make the gyratory velocity less below, and the friction of one air layer on another is very slight aloft, then the less the altitude of the air disturbances above the ground, the more powerful will be the effect of the upper air current, and the nearer to the center will be the region where the air circulation changes from the cyclonic to the anticyclonic. This is analogous to the general hemispherical circulation of the atmosphere. The directions of these air currents are shown in Fig. 65. The full arrows indicate directions of lower currents, and the broken arrows those of upper currents.

It is probable that some cyclones are due entirely to the breaking-up into vortices of steady air currents, while others are due entirely to the local unequal heating of the air, and still others are due to a combination of the two processes.

Moisture most probably plays a very important part in the maintenance of cyclones, since the large quantities of heat liberated by condensation must retard the normal cooling of ascending air very appreciably, and thus continue the condition of instability. Aside from the fact that local unequal heating of the air may arise from the freeing of heat by condensation when upward currents of moist air are caused by outside influences, there is very great uncertainty as regards the influence of moisture in the formation of cyclones.

Air Density. — The local density of the air is dependent mainly on its temperature, and to a slighter degree on its humidity. It has been shown that the lines connecting regions having air of equal densities (over Europe at least) run parallel to the coast, and that the density decreases towards the interior of the continent in summer, and towards the ocean in winter. The study of air densities in connection with cyclones and anticyclones has not been thoroughly carried out; but it has been stated that when the closeness of the lines of equal density signify an unstable condition of equilibrium, then cyclonic disturbances tend to form, and they move along parallel to the lines of equal density. It is very probable that the accurate and careful study of the densities of the air will lead to a better knowledge of the formation and behavior of secondary or local cyclonic disturbances.

ANTICYCLONES.

Classes of Anticyclones. — There are several classes of areas of high barometric pressure. One class is due to the heaping-up of the air thrown off by centrifugal force from

the cyclonal whirls. Such is the ring of high pressure which encircles the globe at about latitude 30°. When cyclonal areas enter this ring of high pressure from without, the air is still further heaped up on their outskirts; and these masses of air are forced out from the main ring especially on its poleward side, and are made to project far into the higher latitudes between the successive cyclones.

Another class of high-pressure areas is due to local cooling of the air. These are the isolated areas of high barometric pressure, which are essentially phenomena of the colder and continental latitudes, where the lower air is cold and dense. The isobaric surfaces being thus brought close together at the cold center, the upper air flows down them towards the center from the warmer regions without, and increases the pressure at the center.

Cyclones and Anticyclones contrasted. — Anticyclones are by no means so well marked in their limits and characteristics as cyclones. In anticyclones the air pressure is greatest at the center, and decreases thence in all directions somewhat irregularly; and on the side where a cyclone is to be found, this decrease continues practically uninterruptedly to the region of lowest air pressure in the cyclone.

Anticyclones are usually an accompaniment of cyclones; and in fact where a cyclone exists, with its deficiency of air, there must lie at not a great distance from it one or more anticyclonic areas into which the air has overflowed from the upper region of the cyclone. Sometimes the cyclones are surrounded by a ring of anticyclones, but the continuity of the latter is nearly always broken by slight depressions or troughs in which the air pressure is lower than the maximum barometric pressure in the anticyclones, which in most cases is above 30 inches.

Air Pressure in Anticyclones. — In ordinary anticyclones of our middle latitudes the barometric pressure frequently reaches 30.6 inches, and in some cases it exceeds even 31 inches. The isobars are even more irregular than in cyclones, and are usually elongated (elliptical), with the greatest axis pointing most frequently about northeast and southwest, although it may be directed towards any point of the compass. The longest diameter is on the average about twice the length of the shorter, but sometimes exceeds it fourfold. The relative length and direction of the axes depend a good deal on the adjacent areas of low barometer. When an anticyclone is situated between two well-developed cyclones not far apart, then the anticyclone becomes much elongated; and when this latter condition obtains there are sometimes two or three distinct centers, which differ in maximum pressure by only slight amounts.

The dimensions of anticyclones exceed those of cyclones. It has been found that the diameter of those in the winter time, with extraordinarily high pressures, was from 3,000 to 4,000 miles; and the preceding and following cyclone centers were distant fully 2,000 miles from the center of the anticyclone, and the preceding cyclone was much the deeper of the two, and the steepest gradients were on this side. Such relations are shown in Fig. 74.

Moisture in Anticyclones. — Anticyclones are characterized by a dry cool air and little cloud, and but slight precipitation. The descending air at the center tends to dissipate the cloudiness. This is due to the fact that as the air descends it is subject to a greater compression, becomes denser, and the temperature is raised; by which means the relative humidity is decreased, and cloud is dissipated. Ground fog frequently occurs in anticyclones, because on unclouded nights just at and very close to the ground the

temperature abruptly decreases to below the point of condensation, and fog is formed.

Temperature Distribution in Anticyclones. — Near the earth's surface the temperatures are generally below the normal air temperatures; and usually the more pronounced the anticyclone (i.e., the greater the air pressure), the lower the temperature. At higher altitudes, however, the air is but little colder than near the ground, and in many cases there is even an inversion of the usual temperature decrease with altitude, and it is actually warmer above than below. This is due to the excessive radiation of heat from the ground into the clear sky. In the anticyclones of our middle latitudes the air is colder on the eastern side.

Direction and Velocity of the Wind in Anticyclones. — Near the surface of the ground in anticyclones the air motion is directed spirally outward. The winds at high altitudes are, on the contrary, directed spirally inward; and at the center of high pressure there is a vertical downward current connecting the inflow above with the outflow below.

The wind velocities in anticyclones are much smaller than those in cyclones. The inner downward current has probably very much less velocity than the upward current in cyclones, and usually amounts to but little more than a settling-down of the air towards the earth's surface. Calms are very frequent within anticyclones.

The Direction and Velocity of Propagation of Anticyclones. — Anticyclones, like cyclones, drift along in the general direction of the great air currents. Since they occur mostly in the middle and higher latitudes, their direction is generally easterly. In the United States the anticyclones move, on the average, in a southeasterly direction, that is, at right angles to the direction of greatest elongation; but in winter the direction is a little more east-

erly. The velocity of translation is not so great as that for cyclones, nor are the paths pursued so long. The reason of this is not well understood, but it is supposed that some process in the formation or continuance of cyclones accelerates their movement in certain directions; and this feature is lacking in the anticyclones, which lag behind.

Excessive High Pressure in Anticyclones. — In the case of an ordinary anticyclone at the rear of a cyclone, we have very little cloud, and consequently the outward radiation is very rapid and great, and the ground and lower air layers are still further cooled. This causes the lower air to become still denser, and the isobaric surfaces are lowered so that the air at greater altitudes in the surrounding region flows down these surfaces and in towards the center. This increases the air pressure, which was already high, and causes it to become very high if the air becomes sufficiently cooled. This accounts for the extremely high air pressures observed in winter.

In considering both cyclones and anticyclones, it must be remembered that their lateral extent is perhaps a thousand times their thickness, and the mass of gyrating air may be regarded as a relatively thin disk. There is thus a chance for a great waste of force through friction, and innumerable opportunities for the interference with the horizontal circulation by means of vertical currents. The original forces which give rise to these phenomena can thus soon become used up.

Relation of Cyclones to Anticyclones. — Anticyclones are a natural consequence of cyclones, because, when there is a diminution of pressure in one place (in the cyclone), there must be a corresponding increase in another place (in the anticyclone). The air movement between a cyclone and an anticyclone is shown in the accompanying diagram

(Fig. 73), where, for the lower layers of the air, the barometric minimum or cyclone is shown on the left, and the barometric maximum or anticyclone on the right; and the wind directions there are shown by the full-drawn arrows. At higher altitudes, however, in the cloud region, the wind direction for which is shown by the broken-lined arrows, above the cyclone the direction of motion of the air is seen to be anticyclonal and away from the center, while above the anticyclone it is cyclonal and towards the center. Thus the upper current from the cyclone replenishes the air lost from the anticyclone at the surface, and the lower current from the anticyclone replaces the air lost from above the cyclone.

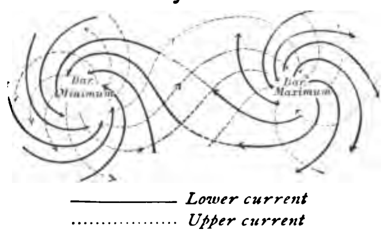


FIG. 73.—INTERCHANGE OF AIR BETWEEN CYCLONE AND ANTICYCLONE (NORTHERN HEMISPHERE).

The cyclones coming from the tropics have to break through the ring of high pressure at about latitude 30°; but there the absolute mass of air is so great, and the cyclone so limited in extent, that the air thrown out by them does not cause much of an increase in the pressure; so that, while there are well-marked cyclones in the lower latitudes, yet the anticyclones with excessively high pressure are not to be found there.

The simultaneous distribution of cyclones and anticyclones over the northern hemisphere is shown in Fig. 74, which gives the isobars on Dec. 15, 1882. This shows an area of high pressure (an anticyclone) over North America, and one over Asia, with areas of low pressure (cyclones) over the eastern Pacific Ocean and the North Atlantic Ocean. A less pronounced anticyclone is also to be seen over the southern part of the North Atlantic Ocean.

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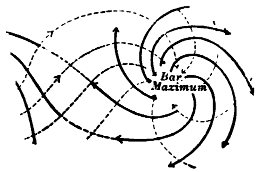

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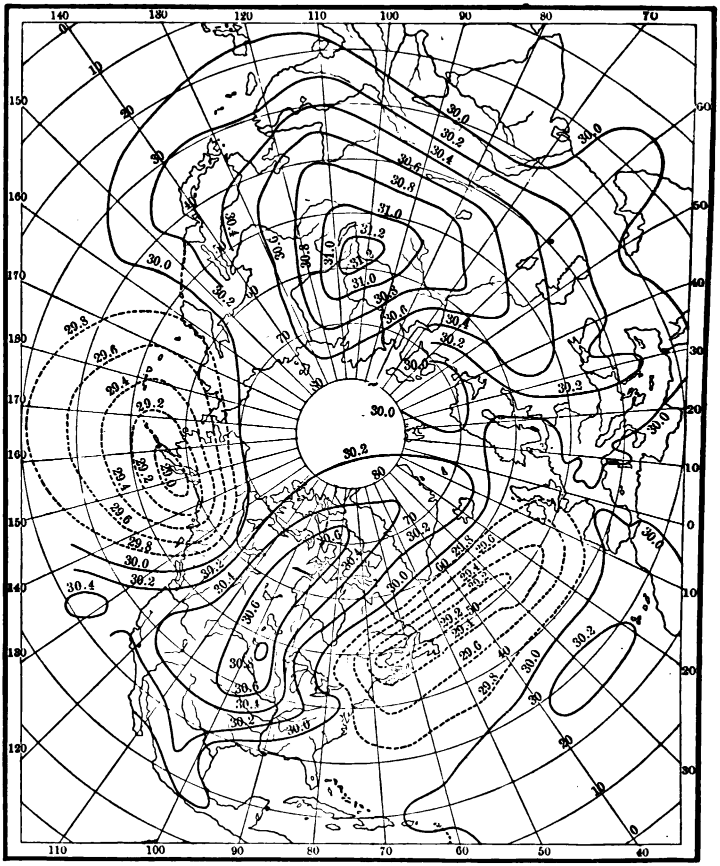


FIG. 74. — CHART SHOWING ISOBARS, DEC. 15, 1882 (AFTER LOOMIS).

Permanent Cyclones and Anticyclones. — In the charts showing the average air pressures over the globe, it is seen that there are local regions of high and of low pressure, — the so-called permanent anticyclones and cyclones.

CHAPTER X.

LOCAL AND MISCELLANEOUS WINDS.

Irregular Local Atmospheric Disturbances. — In those regions where there arise disturbances of the condition of equilibrium of the atmosphere over a limited area, there is found a class of phenomena of which the thunder squalls, spout phenomena (such as waterspouts, sand spouts, and tornadoes), and the straight blows (*derechos*) form the principal members.

These are all somewhat closely related as to their cause; and the final development of one form or another is due to variations in the conditions under which they arise, and somewhat also to the geographical locations. While these phenomena are individually of but small extent, yet the conditions favorable to their formation may extend over a considerable area, so that, instead of individuals, groups may exist at any one time, and these progress in parallel tracks.

TORNADOES.

A Tornado is a progressive, limited, local, violent whirlwind, characterized by a funnel-like cloud which hangs suspended from an intensely black mass of storm clouds; the apex of the funnel cloud sweeps over the earth's surface, sometimes touching it, and sometimes receding from it, to come down again to the ground farther on in the course of the cloud as it moves forward. An ideal picture of a tornado in the United States, as viewed when looking toward the east, is shown in Fig. 75. The peculiar funnel-like cloud may be seen at the center, receding in a northeasterly direction from the scene of destruction.

The first visible precursor of one of these tornadoes is a heavy bank of clouds appearing in the southwest, and later in the west and northwest; in these a violent commotion is observable, with adjacent clouds rushing in from the southeast, east, and northeast, towards the center of the disturbance. If the clouds are light in color, they resemble smoke clouds from a large fire; if they are dark, they are of a peculiar greenish hue, which increases in intensity as the

FIG. 75. — PASSAGE OF A TORNADO (THE OBSERVER FACING THE EAST).

clouds approach. Sometimes these dark clouds look like the dense volumes of smoke emitted from the smokestack of an engine. It is among such clouds that the first appearance of the tornado funnel cloud is to be noticed. It then seems to descend toward the earth by gradual lengthening or growth of the funnel. The near approach of the cloud is announced by a heavy rumbling sound like that of an approaching railroad train, or of distant thunder,

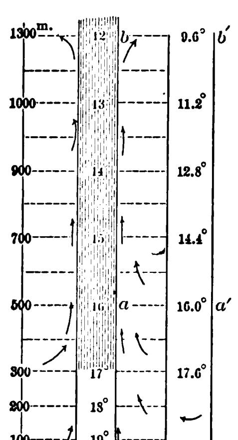
The noise increases as the cloud draws nearer, and its passage creates such an uproar that all other sounds are deadened. The observer in front of this rapidly moving mass of clouds finds himself in the region of a gentle south breeze of warm air, or of calm air and oppressive heat. This quiet condition is suddenly disturbed by the tempestuous whirlwind caused by the tornado itself, which cuts a swath in its narrow track, felling everything before it in an irresistible manner. The passage of the funnel-shaped cloud is succeeded by a sudden lowering of the temperature, and by a calm or gentle breeze.

Air Circulation in Tornadoes. — There is a horizontal current of warm moist air flowing in from all sides (but not from a great distance) towards a center, and an upward current at this center which is fed by the horizontal current just mentioned. Above these there is an outflow away from the center. The motion of the upward current is not directly upwards, but takes a spiral form; and its position is shown by the outlines of the peculiar tornado cloud which hangs in funnel shape, suspended from the intensely black mass of storm clouds (see Fig. 75). The direction of this spiral rotation has been observed to be opposite to that of the hands of a watch, in the cases where this phase has been noticed, but it has not been possible to obtain frequent careful observations of it.

Formation of Tornadoes. — Tornadoes are caused by local differences of temperature. The air having become abnormally heated over a central area, there results a difference in pressure between the air of the inner region and that surrounding it; from this there arises a flow of air spirally inward towards the center, and as it is approached, the velocity is increased. The principal condition for the formation of a tornado is the local unstable condition of

the air, due to the abnormal heating of a mass of air either at the earth's surface or at some locality above it. This mass of air, being warmer than the surrounding air at the same level, is in unstable equilibrium; and when some slight disturbance frees it from its abnormal position, it is forced upwards, by the pressure of the air below and around it, to that altitude where its temperature and density are normal (that is, the same as the temperature and density of the surrounding air); there the power causing the vertical current ceases.

In this ascent the air cools by expansion; but in ascending moist air, when condensation occurs the cooling is slower, and consequently the difference in temperature between the outside and inside column of air becomes greater (Fig. 76), and the upward velocity is increased. But this ascending air at first accumulates above, and the pressure thereby increases, so that at some altitude the air pressure over the ascending column becomes greater than in the surrounding air. The difference between the interior and exterior temperature increases up to this level; but above this point it decreases (owing to the greater density of the air above it), and at some higher altitude the temperature is the same without and within the current, and still above this (and also below the altitude of condensation) the decrease of temperature with the ascent is more rapid within than outside. At the altitude where the temperature becomes the same inside and outside of the current, the air pressure also becomes the same. Below this level the temperature is greater and the pressure less in the ascending current with condensation than outside of it, and above this level the condition is reversed.



Altitudes in meters; temperatures in C.°

FIG. 76. — VERTICAL DECREASE OF TEMPERATURE (SPRUNG).

This increased pressure above forces the air out on all sides as it moves upward, and this air flowing over into the adjacent air mass makes it heavier, and the air below is forced in from it to replace the air which has moved upwards. There is thus caused a vertical air circulation upwards within the central area, and downwards on all sides around it; while an outward horizontal current above, and an inward horizontal current below, connect the two vertical currents. This circulation may take place when the unstable condition exists for any air mass, whether this condition be near the earth or at high altitudes, or whether it extend through the whole vertical air column or not. If this circulation took place without any rotary motion, the velocities would be small, and the flow of air gentle; but because of rotation the velocities increase, and near the center become very great, according to the law of conservation of areas. This rapid gyratory motion near the center gives the tornado winds their tremendous velocities.

With the increasing wind velocities towards the center there is an increasing centrifugal force, and the air is forced away from the center as in the case of cyclones, and thus gradients arise in the air pressure; and these gradients are steeper the nearer the center is approached, so that there is a decrease in the air pressure towards the center of the whirl, and near the center this decrease becomes relatively very great in the case of the great rotational velocities met with in tornadoes.

Sustaining of Tornadic Action. — The difference in temperature within and without the central upward current of the tornado, which is necessary for its continuance, could not persist long enough for its full development if it were not for the freeing of latent heat through condensation. This latent heat becomes sensible, and retards the cooling of the inner air column. It is therefore when the air is saturated with moisture that the conditions are best for the maturing of a tornado; and when condensation no longer takes place, the tornado action must soon cease.

When and Where Tornadoes Occur. — The conditions necessary for the formation of tornadoes are to be met

with at low altitudes in the moist atmosphere of the lower and middle latitudes. They occur in western Africa,[1] southern Asia, occasionally in the lower middle latitudes of Europe, but most frequently in the central and eastern portions of the United States. Most of our knowledge of these phenomena is obtained from the study of those occurring in the last-named country, and the following remarks apply to them. The true tornado is probably not observed above the 50th parallel of latitude. Tornadoes in the United States form most frequently several hundred miles (300 to 500) to the southeast of the center of a cyclone which has a long troughlike central depression extending north and south or northeast and southwest. This region is the one of greatest heat and moisture, and of surface winds of considerable force from the south. The regions of greatest frequency of tornadoes are the lower Missouri, the central Mississippi, and Ohio river valleys; and the season of their greatest frequency is the late spring and summer for these regions, but farther south (in Georgia, for instance) many tornadoes occur in the early spring. Few tornadoes occur in the United States west of the 100th meridian. The hours of greatest frequency of tornadoes are from 3.30 to 5 P.M.

Paths of Tornadoes. — The path of destruction which marks the central region of a tornado varies from a few feet up to 2 miles, the average being about a quarter of a mile. The length of such tracks, which mark the places where the tornadoes reached to the earth's surface, vary from 300 yards to 200 miles, with an average of about 25 miles. The funnel-shaped cloud at the earth's surface

[1] Where they occur in the southern hemisphere, the directions of motion are the reverse of those in the northern hemisphere, because in the former the cyclonal rotation is with the hands of a watch.

varies in diameter from a few yards to an unmeasured limit, but the average is probably several hundred feet.

Velocity of Motion of Translation of Tornadoes. — The velocity of progression of tornadoes varies from 7 to 100 miles per hour, with an average of about 44 miles per hour. The same tornado cloud may sometimes remain almost stationary for a while, and then dart forwards with great velocity. The average time occupied in passing a point is a little over a minute.

The direction of motion of translation of the tornado is nearly always from southwest to northeast.

Air Pressures in Tornadoes. — The normal air pressure is about 14.7 pounds per square inch; and if, as is possible, the air pressure is reduced one fourth of this amount at the center of a tornado, then it is lessened there about 3.7 pounds per square inch, or 533 pounds per square foot. Supposing, now, that the tornado passes over a house in which the air is at the normal pressure (that is, on the inner and outer walls there is a pressure of 2,117 pounds per square foot), then the pressure on the outer walls during the passage of the center of the tornado will be suddenly diminished by 533 pounds per square foot, and there will be an excess of pressure by this amount from within outward. The diminution of the outside pressure comes very suddenly, and this inside pressure acts like an explosion. It is not to be wondered at, then, that doors and windows are blown out, and walls even torn down, when this great force so acts suddenly on them in an outward direction.

Destructive Winds in Tornadoes. — The enormous wind velocities in tornadoes are measured only indirectly by the mass or strength of objects which they have moved or broken. It is probable that they reach or even ex-

ceed 400 or 500 miles per hour in some cases. The winds are greatest near the center, and decrease from thence outwards; they are less in front than at the rear. Tornadoes are usually accompanied by hail fall, and very frequently by manifestations of atmospheric electricity.

On the right side of the tornado the winds are stronger than on the left-hand side of its direction of progression; and the path of destruction extends much farther from the center on the right than on the left side.

The winds on the left (or northwest) side are the weakest, because, for one thing, the direction of progression is there directly opposed to that of the gyratory motion. In Fig. 77 the long arrow shows the direction of progression, and the small arrows the gyratory motion; and it is seen that on the right-hand side the two motions are in the same general direction, while on the left the directions are opposed to each other.

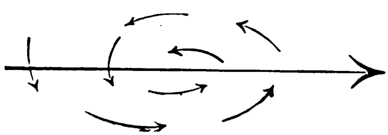


FIG. 77.

The wind force in tornadoes depends directly on the velocity of the air current and the density of the air. It has been computed, that, in the case of a tornado in which the gyratory velocity was 10 feet per second at a distance of 3,300 feet from the center, the velocity at 70 feet from the center would be 460 feet per second, or 310 miles an hour, which would give a pressure force of about 300 pounds per square foot on any plane surface exposed squarely to the wind. In case the ascending current in a tornado had a velocity of 100 miles an hour at the altitude where the air pressure is 15 inches, this would prevent a hailstone 2.5 inches in diameter from falling to the ground. Large raindrops 0.1 of an inch in diameter require an upward current of only 16.5 miles an hour to sustain them when the air pressure is 23 inches; while small drops only 0.003 inch in diameter require an air current of only 3 miles an hour to prevent their falling to the ground.

The size of raindrops may be taken as indicating roughly the velocity of the ascending air current. When the raindrops are fine, there is a weak upward current; but when they are coarse, there is a more powerful one.

Safety in Tornadoes. — Concerning the safety of those who find themselves in the path of a tornado, it may be said that it is best to enter some subterranean vault, such as an ice cellar. The cellar of a house is preferable to the house itself if the house is of wood; but if it is of stone or brick, one should not enter the cellar, but leave the building. The southwest corner of a cellar is the safest.

If one must escape a tornado in an open place, he should first take the bearings of the approaching tornado and its course. If the tornado is discovered to the west or southwest of him, he should move off with dispatch towards the southeast or northwest, respectively. If the tornado is far to the south or to the north, one is usually safe. If one is caught in the whirlwind's blast, it is best to lie prone on the ground in an open field rather than seek shelter in a wood or under trees.

Frequency of Tornadoes. — During the 10 years from 1877 to 1887 the number of tornadoes reported in the United States averaged 146 a year. Observations do not as yet prove that tornadoes are increasing or decreasing in frequency.

Multiple Tornadoes. — Frequently on the same afternoon a number of tornadoes will form. They may be close together or several hundred miles apart, and they move in parallel directions. Thus they form a group or band of isolated whirlwinds, all of which have for their origins the same general conditions.

THUNDERSTORMS.

Thunderstorms and their Attendant Phenomena. — Thunderstorms are local progressive atmospheric disturbances occurring in most latitudes inhabited by mankind, in

regions where there is considerable atmospheric moisture. They receive their name from the electrical phenomena which distinguish them. They are usually accompanied by rainfall. Their attendant phenomena in our latitudes are as follows: First dark clouds are seen lying low in the western sky, and light southern winds are experienced, the air being warm and sultry. Perhaps an hour later the clouds have mounted to near the zenith, but the same relative quietness of the air continues near the ground; and the heat is perhaps not so great, owing to the sheltering influence of the clouds. The latter usually present an appearance something like this: On the front side of the thunderstorm there are grayish white or reddish clouds hanging over and in front of the main rain cloud. Above these, dense dark-gray and violet cumulo-stratus are seen, as also dome-like cumulus clouds, which are separated from the cumulo-stratus. Often these are interspersed with one or more thick cumulo-stratus-like cloud layers, and above all is the widely distributed cover of cirro-stratus.

The first thunder is heard before the cloud reaches the zenith, and the first rain commences after it. The interval between the first thunder and the beginning of the rain varies from a few minutes to perhaps half an hour or more. Just before (usually only five minutes or so) the rain begins, there comes from the west or northwest a brisk wind, which suddenly increases in violence, and becomes a squall, which is the name given to a sudden violent and not long-continued wind; this, however, dies down very much, after the rain begins. The time of heaviest rainfall varies: sometimes it occurs at the beginning, and sometimes in the latter part, of the time when the rain falls. The thunder is loudest, and lightning strokes occur some minutes after the rain begins.

Soon after this the western horizon loses its dark aspect, and begins to lighten up a little, and finally the clouds break there, and blue sky is seen. The storm clouds then pass by overhead, and the rain ceases some minutes before their western edge reaches the zenith. Although the temperature may have fallen even as much as 25° F. during the storm, yet it frequently regains its former condition after the storm has passed. Rainbows appear in the latter part of the rain period of the storm. The last thunder is usually heard after the rear edge of the clouds has passed the zenith. Individual thunderstorms seldom last (between the first and last thunder) over two hours, but frequently one thunderstorm follows another in such quick succession as to appear to be part of the same storm. In our latitudes the usual direction of translation of thunderstorms is easterly.

Classes of Thunderstorms. — Thunderstorms are now generally divided into three classes, — *cyclonic*, *heat*, and *winter* thunderstorms.

Cyclonic thunderstorms accompany the well-developed areas of low barometric pressure, and they have a circulation somewhat analogous to that of cyclones, becoming in extreme cases tornadoes.

Heat thunderstorms are the result of the local heating of the lower air, which makes its condition unstable. They need for their development a moist quiet air warmed by the sun's rays, and they occur in their most distinct state outside of the regions of strong ascending and descending currents of cyclonic and anticyclonic areas.

Probably many, if not the majority, of our thunderstorms are a combination of these two forms.

Winter thunderstorms occur most frequently at night, especially in high latitudes, and they are more frequent

near the coast than inland. These winter storms are of more limited area, travel faster, are shorter in duration, and their lightning is more destructive, than the summer thunderstorms.

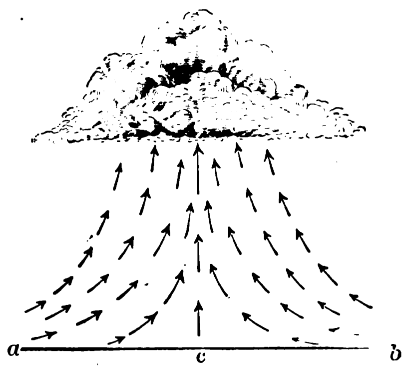


FIG. 78. — UPWARD AIR CURRENT IN THUNDERSTORM (AFTER FERREL).

Heat or Stationary Thunderstorms. — In a stationary thunderstorm, that is, one having no or but very slight progressive motion, we have at first an ascending current of moist warm air over some limited region. Air then flows in from all sides, and a vertical circulation arises (Fig. 78). When the ascending air reaches an altitude where its dew-point occurs, then condensation begins, and a cloud is formed which has a flat base. Above this, precipitation occurs; and the rain falling through the air beneath cools it, so that it contracts, becoming heavier than the surrounding air, and, being also pressed downward by the falling rain, falls downward, after overcoming the gentle upward current which already existed there. The movement of the air downward causes an outflow of the air below it, and, forcing its way into the origi-

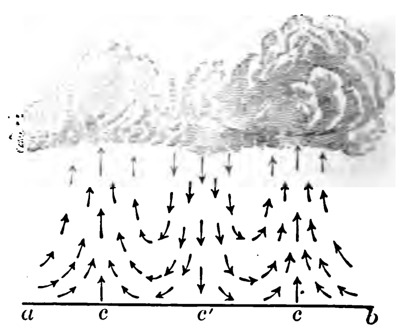


FIG. 79. — AIR CURRENTS IN THUNDERSTORM (AFTER FERREL).

nal ascending current which surrounds it, causes it to spread out, thus increasing the size of the atmospheric disturbance and the cloud region overhead. The air thus forced out from below at the center also takes part in the vertical circulation of the surrounding ascending air (Fig. 79). Since this occurs on all sides, the outer upward currents are arranged symmetrically around the inner downward current.

There is, then, a central region of greatest air pressure caused by this central cooling and downward flow of air, which reaches its maximum at the earth's surface; and outward from this the temperature increases and the pressure decreases, and this change is very rapid within a short horizontal distance at the dividing line between the descending and ascending currents. It is in this region of steep gradients that the squall or sudden wind occurs. The air which descends at the interior, forces its way out under the ascending current encircling it, and, in part at least, very soon joins the latter in its upward movement, after its horizontal force is somewhat spent. By this mixture of the cool interior air with the warmer outer air, the latter becomes cool on the inner side, and changes there to a descending current, so that the ring of steep gradient and squalls is enlarged. This process lasts until the stable equilibrium is restored, when the storm disappears.

Cyclonic or Progressive Thunderstorms. — It must be rare, at least in middle latitudes, that the air is quiet enough to permit the formation of a symmetrical thunderstorm such as has just been mentioned. In the southern and southeastern parts of the large cyclones, and at a distance of several hundred miles from the center, where our thunderstorms usually occur, there is an easterly motion (the wind varying from northwest to southwest) of the air. It is this easterly motion which carries the thunderstorm on-

wards; and since the motion is greater above than below, new masses of air are introduced on the front side near the earth's surface, and thus a fresh supply of moisture is drawn into the vertical currents to keep up the precipitation; and on the rear side this air is left drier and cooler. Not only does the storm progress in this easterly direction with the upper air currents, but it increases also on the side of greatest moisture and heat, which in this case is the preceding or front side.

The local gradient which occurs between the air of the inner and that of the outer regions, and which is indicated by the sudden rise of barometric pressure at the central region in the thunderstorm, is sufficient to cause a wind as powerful as that observed at the front edge of the storm.

Thunderstorms seem to occur under much the same conditions as those favorable for the formation of tornadoes. The latter are nearly always accompanied by lightning; and where the tornado is too far distant to render its distinctive funnel cloud visible, but not too far for its thunder to be heard or its lightning seen, it is regarded as a thunderstorm.

Progressive Movement of Thunderstorms. — Although the thunderstorm thus commences small in size, it does not remain so. As the storm progresses (moving generally in a direction somewhat easterly), it broadens out so that it presents an ever-increasing length of front, until it is dissipated or else breaks up into smaller storms. This is illustrated in the following diagram (Fig. 80), in which the relative length of the front at successive hours is shown.

Sometimes thunderstorms expand in all directions from a central point, but this occurs usually when the storm is very local and stationary, or in a stationary stage. The direction of propagation of thunderstorms is usually that of

the local wind belonging to the cyclonic system in which it appears. In Europe, thunderstorms move with an average velocity of from 22 to 25 miles per hour; and in general the velocity is probably nearly equal to that of tornadoes.

Time of Occurrence of Thunderstorms. — Our thunderstorms occur most frequently in June, July, and August, and least frequently in winter; they occur most frequently in the middle of the afternoon, and least frequently in the early morning hours. In some places it has been found that more thunderstorms occur one or two hours after midnight than in the hours just preceding and following.

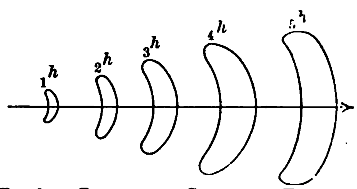


FIG. 80. — PROGRESSIVE GROWTH OF THUNDERSTORMS.

Variation of Number with Latitude. — Thunderstorms are in general most frequent near the equator, and decrease with increase of latitude; but their number depends also on the temperature, the moisture, and the movement of the air. For instance, in the tropics, during the season of calms, thunderstorms are frequent; but during the season of trade winds they are rare, because, the air as a whole being then in constant motion, the initial condition of unstable equilibrium necessary for the formation of thunderstorms seldom occurs.

Changes in Meteorological Elements during Thunderstorms. — In the passage of thunderstorms the meteorological elements undergo the following changes: Before the thunderstorm the air pressure and the relative humidity decrease, and the temperature rises, the wind being generally weak; so that at the beginning of the storm the pressure and relative humidity are at their lowest, and the temperature at its highest point. At the moment of the bursting of

the storm the air pressure and relative humidity increase very rapidly, and the temperature falls; the wind also suddenly becomes strong, and sometimes it as suddenly subsides almost immediately afterwards, while at other times it increases until near the close of the thunderstorm. Towards or at the end of the thunderstorm, the air pressure

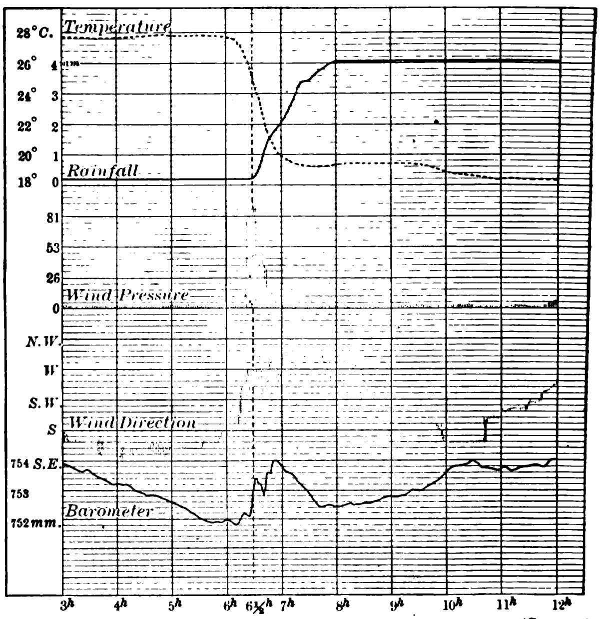


FIG. 81.—METEOROLOGICAL ELEMENTS DURING A THUNDERSTORM (SPRUNG).

and relative humidity reach their maximum, and the temperature its minimum. These relations are shown by the preceding diagram (Fig. 81), which shows the variation of the meteorological elements, at a place in the path of a thunderstorm, during the afternoon and evening of the day on which a thunderstorm occurred at half-past six o'clock.

The rain areas in progressive thunderstorms increase with the spreading-out of the upward air currents and gradual growth of the storm; and at any one time the rain area has an average form as shown in Fig. 82. The thunderstorm is carried along by the wind which is least at the surface of the earth; so that the warm moist air on the front side feeds the air in the thunderstorm like a lower current in a westerly direction, and the storm grows in extent towards this lower feeding current, which supplies it with moisture. In the rear the storm dies out from lack of moisture to sustain it; also the storm spreads toward the south when warm moist southern winds blow into it, and with the diminution of this moisture by precipitation there is a dying-out of the storm. The growth of the storm is therefore toward the east and south. On the north side the storm increases in size but slowly and in a more regular form, and is even wider than on the south side, because on the north the changes of moisture from one stage to another take place less rapidly, and the moisture is held in the air longer.

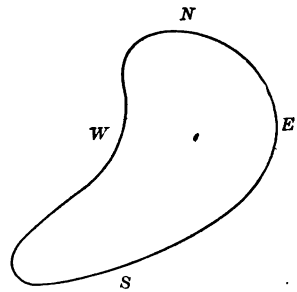


FIG. 82.—FORM OF RAIN AREA IN A THUNDERSTORM.

Ideal Sketch of a Thunderstorm.—The following diagram (Fig. 83) gives a detailed sketch, in vertical section, of the clouds and their circulation in a thunderstorm, or in a rain squall not accompanied by thunder.

On the right-hand edge or front there is seen the outflow of air below, with a gentler inflow towards the cloud above. In the rainy region, shown by the shading, there is a downward flow of air, and to the rear of this is the upward current already mentioned. The

lower edge of the roll-like clouds at *b* and (*b*) may be from 1,000 or 1,200 to 1,500 feet above the ground, while the dome-like vault of the cloud in the rain section is perhaps 2,500 feet. At (*b*) there may be almost or quite a cessation of the rain, and then it may commence again, and perhaps another cessation may occur as shown. The cloud

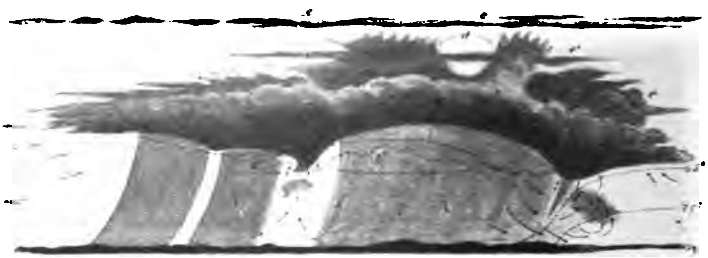


FIG. 83. — CLOUDS AND WINDS IN A THUNDER SQUALL WHICH IS MOVING TOWARDS THE RIGHT (VERTICAL SECTION). (AFTER KÖPPEN.)

forms are distinguished as follows: At *c* are the cumulus cloud heads or tops projecting from the mass of dark storm cloud *a*. At *b* and (*b*) are dense rolls of clouds hanging down lower than the main mass. At *ee* there is a continuous cirro-stratus layer; and lower, at *e'*, is another

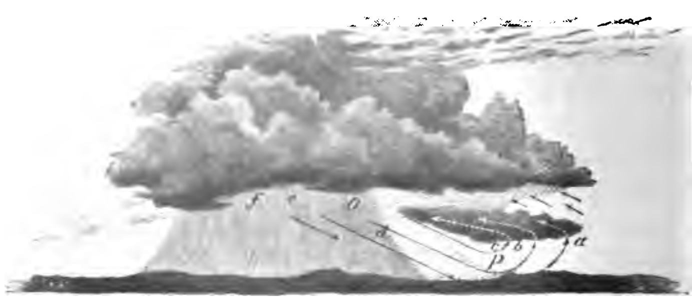


FIG. 84. — CLOUDS AND WINDS IN A HAIL SQUALL.

layer partly broken, and pierced by the heads of the cumulus clouds below. At *d* instead of rounded tops there are cloud tops of fantastic forms resulting from the outstreaming of the air. The downward current between *b* and (*b*) is due to the heaping-up of air at *a*, and replaces

the air carried upward on either side at *b'* and (*b*). At the rear of the storm are frequently seen, as at *g*, isolated cloudlets due to the local complete saturation of the cold air just after the storm has passed.

In case of hail formation and hail fall, the frozen rain will fall in the region at the front side shown by the coarser dotted shading in Fig. 84.

Hailstones may form in the upward current of moist air, and then drop out to one side of the upward current down into the region of rain, where they will have water added to their exteriors; and then they may be carried up, in the ascending current into which they have fallen, into the region of freezing. The hailstones may become largely increased in size by repetition of this process. Sometimes the central kernel of a hailstone is a snowflake, which at first becomes thoroughly wet by contact with rain, and then is carried up by an air current to the region of freezing; and then, following the process just described for a hailstone, it may increase to a very large size. A variety of such processes must occur in order to account for the different kinds of hailstones which fall during storms of this kind.

Thunderstorms, tornadoes, and hailstorms have a common relation to cyclones as far as position is concerned, since they all depend on the same condition of great humidity and instability of the atmosphere. In addition to those conditions necessary for the formation of a thunder squall, tornadoes require the local conditions which cause the gyratory motions peculiar to them; and hailstorms require the powerful ascending current which extends up to altitudes where freezing can take place.

Spouts.

Spouts are cloud-like or fog-like phenomena of a slender, more or less tapering form, like a long trunk or funnel, formed in the air under certain atmospheric conditions. The funnel-shaped cloud of a tornado is a spout, but there are also column-like spouts which occur in fair weather, especially over water surfaces, and which extend from the earth's surface to the clouds. These are waterspouts, and are the cloud brought down to the earth by the rapid gyratory motion such as occurs at the center of a tornado.

Spouts are due to the rarefaction of the air caused by the centrifugal force driving some of the air from the center of a rotating column of air in which the motion is spirally inward and upward. As the moisture-laden air enters this central region, where the air pressure is reduced so much as to bring the air temperature down to the dew-point, the vapor is condensed into cloud. It is this cloud of condensed moisture which we see as a dark column. Water is sucked up some distance above the water surface at the foot of the column, but probably not very high up into the air. The gyrations being most violent at certain altitudes above the earth's surface, the central rarefaction is in the beginning greatest up there; and the funnel or spout cloud forms there first, and afterwards extends down to the earth's surface if the gyratory velocities are sufficiently great to diminish the air pressure in the central region by such an amount as to cause condensation of the moisture in the lowest air layer. Frequently spouts of little energy are not visible clear to the earth's surface. Fig. 85 illustrates waterspout phenomena of various degrees of energy, and shows how a water

FIG. 85.—WATERSPOUTS.

surface is stirred up by the whirlwind at the base of the spout.

Dust Whirlwinds are really spouts of feeble energy, in which the air is too dry to permit the cloud formation by the condensation of vapor. The dust, however, is carried inwards and upwards, and is held suspended in the air where the motions are strong enough to sustain it; and the outline of this region is rendered visible by the dust particles.

White Squalls, or fair-weather whirlwinds, occur when the conditions are ripe for spout formation, but there is not enough moisture present to form the cloud throughout the length of the central vortex. Up at a considerable altitude, however, a cloud may be formed which will indicate the location of the top of the vortex; and at the bottom the gyratory inflowing wind may create a disturbance which is especially noticeable over a water surface, and in well-developed cases proves of great danger to ships. These squalls are sometimes called *bull's-eye squalls*, on account of the peculiar appearance of the small isolated cloud which marks the top of the invisible spout at the center of the whirlwind.

Cloud-bursts are sudden and excessive downpours of rain, or rain and hail, which have been carried upward or merely sustained and kept from falling by the ascending air currents, until a large amount has been accumulated aloft, when, by some weakening or breaking-up of the ascending currents, the whole or a part of the accumulation suddenly falls to the ground. Cloud-bursts are of most frequent occurrence in connection with tornadoes, where the immense velocity of the ascending current is favorable to the collection and support of great masses of water.

Periodic Local Winds.

Periodic Local Winds are periodic winds which do not depend on the general or secondary circulation of the atmosphere. They are the *land* and *sea breezes*, *mountain* and *valley winds*, and *monsoons*.

Land and Sea Breezes occur daily on the coasts of large bodies of water. When the sun rises, the land is warmed more rapidly than the water surface; and this heat, being communicated to the air above the land, causes an expansion upwards, and the surfaces of equal air pressure are thus raised higher over the land than over the sea. The air over the land flows down these inclined surfaces (so to speak) towards the sea, and the pressure of the air over the sea is thus increased, and the air pressure over the land correspondingly decreased. The increased pressure over the sea causes a return surface wind to set in toward the land region of deficient pressure. This latter air current commences at some distance out at sea, away from the shore, and moves inward towards the land. This circulation is kept up until the air above the land is no longer warmer than that over the water.

At evening, when the sun goes down, the land cools more rapidly than the water, and the air over the land becomes cooled more rapidly than that over the sea. The result is, that the surfaces of equal air pressure are lowered over the land, and the air above flows from above the sea down these surfaces, causing an excess of pressure over the land, and an under wind from the land to the region of deficient pressure over the sea. This land breeze continues until the temperature and pressure differences are adjusted.

Mountain and Valley Winds occur in mountainous regions in relatively quiet conditions of the air. There is

then, especially in clear weather, an upward motion from the valleys during the day, and a downward motion towards the valleys at night. These winds have some analogy to the land and sea breezes. In the night the cold denser air descends from the mountain tops, which cool very rapidly by radiation, toward the valleys below. These downward currents commence first in the narrow valleys on the mountain side, in which the air is least heated during the day; but the descent soon becomes general, and the air flows down to wider valleys below. In the morning the air over lower levels of the ground becomes heated, and the surfaces of equal air pressure over those levels at some little distance away from the mountain are elevated, but directly on the mountain they are not, because there is no air below to expand upward; so that the air flows down these nearly horizontal surfaces towards the upper parts of the mountain, which it strikes at such an angle as to deflect it upward along the mountain side. In addition to this motion, there is an upward motion along the mountain side, due to the action of the sun's rays in the warming of the ground along the slope and the air immediately above it; so that the upward motion of the air along the surface of the ground is the result of these two motions combined. The accumulation of this air at the top of the mountain is prevented by horizontal currents which exist there.

Miscellaneous Winds.

Under this title we shall mention certain winds which have distinctive features, and have therefore received separate names. Some of them belong to the great primary circulation of the atmosphere, and some to the secondary circulation, and others are of local origin.

Cold and Hot Winds. — When the cyclonic and anticyclonic conditions of the middle latitudes are such as to carry to warmer regions the air which has become excessively cooled by local causes, there arise the cold winds which have received their distinctive names in the various regions in which they occur. When the conditions are such as to carry to colder regions air which has been made excessively warm by local causes, warm winds are experienced. The warm winds usually have a poleward and the cold winds an equatorward movement, but either wind may have a motion in other directions.

Cold Winds. — The cold north and northwest winds of the United States, called in the south *northers* and in the north *blizzards*, the *bora* of the Adriatic, the *mistral* of Mediterranean France, the *buran* of Russia, and the *pampero* of South America from the southwest,[1] are due to currents of cold, dry, dense air which flow from the cooler to the warmer latitudes under the control of the secondary atmospheric disturbances. These cold winds occur usually when the temperature changes are very rapid within short geographical distances, and consequently the cold air need not be blown far to reach regions of greater warmth. They consequently occur in winter, and most frequently in those regions where the temperature gradients are steepest. The extent to which they penetrate the warmer regions depends on their persistency and the velocity as well as the volume of the moving mass of cold air.

When the region from which these northerly (in the northern hemisphere) winds blow is very much elevated, as for instance is very markedly the case along the northern shores of the Adriatic Sea, then the air which has be-

[1] It must be remembered that in the southern hemisphere the polar winds come from the south.

come excessively cooled by radiation on these highlands is partly blown, and partly descends from its own greater density, on to the low lands or sea below; and, notwithstanding the adiabatic warming, it reaches the low levels still much cooler than the surrounding air, which it would not do unless there were steep horizontal temperature gradients in the air. Thus are formed the *bora* of the Adriatic, and the *tramontana negra* or *black norther* of Greece.

Hot Winds.—The hot winds from a southerly direction, —the *sirocco* of the central and western Mediterranean Sea, the *leste* of Madeira, the *kahnisin* of Egypt, the *leveche* of Spain, or our own *hot winds* of the central and eastern United States,—and the *zondas* of the South American pampas from a northerly direction,[1] are likewise due to the influence of the secondary atmospheric disturbances of middle latitudes, and are the reverse of the cold winds above described. The air is blown from the excessively heated regions to the cooler. Sometimes it is dry, as in the case of the *dry sirocco* of Spain and occasionally of Italy and Sicily, and the hot winds of our great plains; and sometimes it is very moist, as in the case of the usual Italian or Sicilian *sirocco*. In the latter case the very dry hot air is blown from the Sahara and the northern coast of Africa across the Mediterranean Sea, where it takes up the moisture to almost the point of saturation; and then it reaches the northern shores of the sea very warm and very moist, which renders the air very oppressive.

Foehn Winds.—The *foehn* winds, so called by the natives in the Alps, where they frequently occur, are warm dry winds or hot waves peculiar to some mountainous regions, and are the result of the following process: The

[1] In the southern hemisphere the equatorial winds come from the north.

warm moist air is blown against the side of a mountain range, and a not necessarily rapid upward air current arises along the slope; and condensation usually takes place either by the cooling of the air coming in contact with the colder mountain top or by adiabatic cooling of the ascending air, or by both. When the moisture has thus been abstracted from the air in its passage over the mountains, the air, which begins its descent on the other side of the mountain range, is much drier; and in this condition its adiabatic increase in temperature as it moves downward is much more rapid than the adiabatic cooling in its ascent, for the latter was retarded by the liberated latent heat when the moisture of the air was condensed; and with the descent the relative humidity decreases. Thus, when it reaches low altitudes, the air becomes very warm and very dry.

The decrease in temperature for saturated ascending air is only about half as much per 100 feet as the increase of temperature for the dry descending air which is subject to dynamic heating by compression. The greater the amount of moisture lost out of the air, and the farther the descent of the air, the hotter it becomes, and the more intense are the *foehn* characteristics.

The degree of intensity of heat in the *foehn* depends on the amount of water lost by condensation high up in the mountains, and on the distance of the descent of the air. If no water were lost out of the air, there would be no *foehn*.

One peculiarity of these *foehn* winds is that the descent of air frequently, or perhaps generally, occurs in isolated or disconnected streams, and thus the resulting excessively heated condition of the air below is not continuous, and is found only at intervals; the intervening air not being much, if any, above a natural temperature. Winds of a *foehn*-like character are found to the leeward in most mountain regions where warm moist winds blow against mountain ranges. The hot *Chinook* winds of

the western part of America at the eastern slope of the Rocky Mountains, and on the plains at their base, are due to this *foehn process*, as this drying and warming of moist winds is called. They are especially frequent near the northern border of the United States, where the excessively moist air from the Pacific Ocean coast of Oregon and Washington is carried inland, and it is probable that they extend southward even to Texas.

A single case, which occurred at Fort Assiniboine, Montana, on Jan. 19, 1892, may be mentioned, in which the temperature rose about 43° F. between 2 A.M. and 2.15 A.M., changing from $-5.5°$ F. to 37.5° F. within these 15 minutes. In some cases the temperature rises 80° F. in the course of 6 or 8 hours.

It is very probable that the *foehn process* takes place in the free atmosphere, where air currents ascend, and some water is lost by rainfall; and then the same air enters a descending current, and is brought down to or near the ground again. This would naturally occur in the circulation between cyclones and anticyclones, where the air ascends in the former, and, after losing some of its moisture, descends in the latter, and reaches its former altitude warmer than when it started. The same reasoning also applies to the case of the rising and falling air in the local thunderstorm and tornado phenomena.

The effect of the hot, dry *foehn* winds is very marked. When they occur in the cold season in regions where snow falls, the snow disappears almost as if by magic, and these warm winds are thus sometimes called *snow eaters.* When they occur in the warm season, vegetation is frequently withered, and growing crops entirely destroyed.

Avalanche Winds are the movements of air masses which are pushed ahead of masses of land or snow as they descend mountain sides in land or snow slides. The mass of air is compressed and becomes locally denser as it is pushed ahead of the moving substance. This air movement is

of such violence that it does great damage even at a distance of many feet from the solid mass of snow or earth moved in the avalanche. In the case of a snow avalanche, the limits of these winds are well shown by the cloud of snow particles which they carry forward. The effects of such winds have been especially noticeable in Switzerland, where trees have been broken off by the wind at a distance of 1,500 feet from the mass of snow, and the particles of snow driven to a distance of more than a mile from it.

CHAPTER XI.

WEATHER AND WEATHER PREDICTIONS.

Main Features. — By the term *weather* we mean the atmospheric conditions as shown by the meteorological elements at a particular time or during a short specified period. Climate is the aggregate of weather conditions, and weather is but a phase of climate. Thus we might speak of the weather at any instant or for a day, season, or even a year, but not for such a long period of years as would give the average conditions. These last with their oscillations would be climate. The degree of heat, the amount of moisture, precipitation, and cloud, and the direction and force of the winds, are the main features of our weather. There are certain combinations of conditions which occur together. Thus, in general, heat and wetness, and cold and dryness, go hand in hand; although in middle and polar latitudes, where the seasonal changes are most marked, in the summer time cold and wetness, and in the winter time cold and dryness, go together.

Absolute and Relative Weather Conditions. — In the different localities on the earth's surface we get used to certain normal or average conditions of weather, and the ordinary or average diurnal changes where such exist; but any accidental deviation from these average conditions and changes attracts our notice.

Thus, when we speak of a hot day, it usually means hot in relation to the average temperature at that hour and

season of the year. A condition which we should term hot at one hour of the day might be cool when measured by that usual at another hour. Thus, in summer a temperature of 80° F. at eight o'clock in the morning would seem hot; but the same temperature at two o'clock in the afternoon would not seem hot. The expressions which we apply to heat and cold are relative in their significance, and we make allowance for the diurnal changes.

This is not so much the case when we speak of the wet or dry weather, for then we have the absolute and distinct dividing line between precipitation and no precipitation. The dampness of the air without precipitation is estimated on a scale reaching from extreme dryness with a clear sky to the dense cloudiness of fog.

The wind ranging from a calm to a storm is a prominent feature of the weather; and it is partly noticeable in its direct effects, as when it moves branches of trees, and partly in its indirect effects, as when it causes rapid evaporation or blows the cold air through our clothing, and makes the cold more readily felt.

In going from a region with one climate to that with another, or, as we say, from one climate to another, one must become accustomed to the average conditions in the new climate before he can judge of the weather and its changes by the same standard as that customarily applied in that region.

Seasonal Weather. — During long periods of time which take account of average conditions only, the weather conditions are seasonal; and although there may occur during any one season marked departures from the averages of many seasons, yet these irregularities are small as compared with those which occur during short intervals of time.

If during some particular season the average temperature is, say, only 5° F. higher, or the rainfall a few inches greater, than usual, then we speak of it as being an exceptionally warm or wet season. The study of these long-period weather conditions belongs more properly to climatology.

Current Weather Conditions are studied in a very different manner from the average conditions just mentioned. The true cause of the peculiar weather conditions at a particular time at some one place, and the relation of these conditions to those existing at the same time at adjacent or even quite distant places, have been found out by writing on maps of extensive regions the meteorological conditions existing simultaneously at a great number of places included on the maps. Such maps are called *weather maps;* and they not only permit us to see what kind of weather actually existed at the time the observations were made, but, as we shall see, they also enable us to say with a considerable degree of certainty what the weather has been for a day or two before, and what it will be for a day or two after, that time; and this last we call *weather predicting.*

In the present chapter we shall mainly consider these weather maps and the methods of using them for predicting the weather, or for *weather forecasting;* and we shall confine ourselves to the consideration of the middle latitudes of the northern hemisphere, as these are the regions in which we are most interested, and where such studies are mainly carried on.

It has been found that our weather in these latitudes is mainly conditioned by the phenomena attending the passage of the cyclones and anticyclones moving in the general direction from west towards the east, or sometimes from the south towards the northeast, as already men-

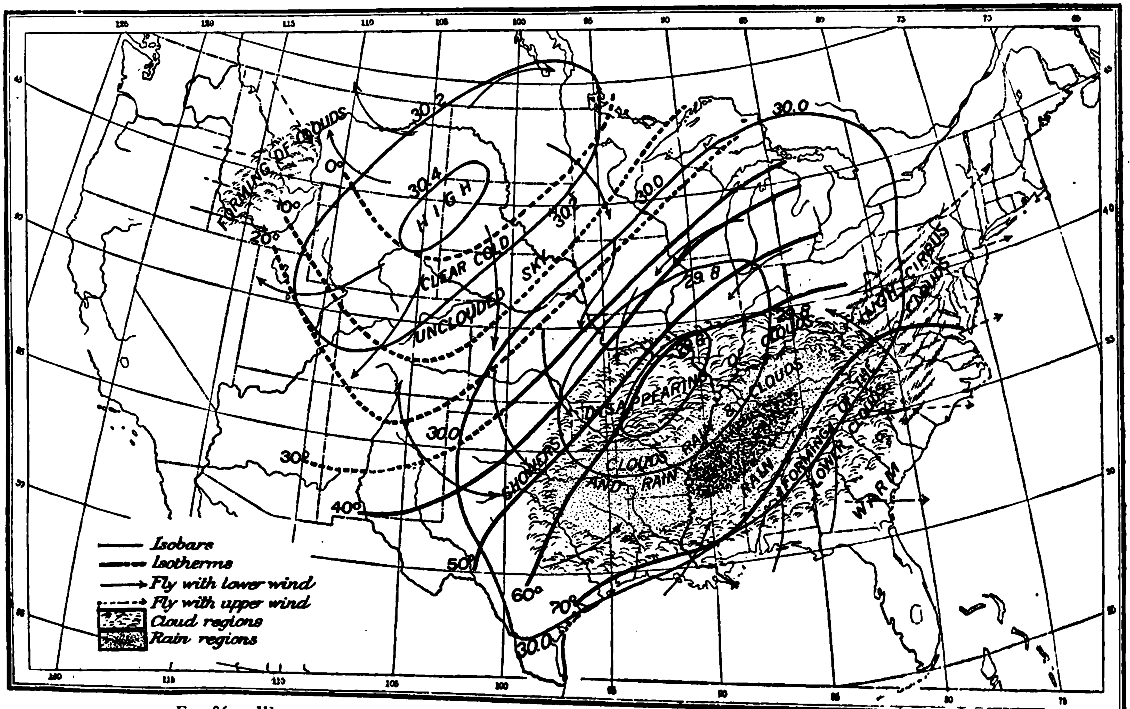


FIG. 86. — WEATHER CONDITIONS FOR AVERAGE AREAS OF HIGH AND LOW BAROMETRIC PRESSURE.

tioned. These phenomena, and their relation to cyclonic and anticyclonic areas, have already been described in preceding chapters, where the distribution of the meteorological elements in cyclones and anticyclones is given.

Weather Conditions in a Cyclone and an Anticyclone. — Fig. 86 shows the combined weather conditions of a cyclonic and an anticyclonic area. On the front (eastern) side of the cyclone we see at first high cirrus clouds moving toward the east or northeast, and towards the center lower clouds moving from the east or southeast, with increased cloudiness and rain; the air pressure diminishes, but the temperature does not change much. Near the center of the cyclone the rain ceases, and cloudiness diminishes.

On the rear (western) side of the center there is some cloudiness and perhaps showers, and proceeding from the center the temperature falls, the air pressure increases, and the wind blows from the west and northwest.

At the south side of the cyclone, proceeding from the front (east) to the rear (west), the winds veer around from southeast through south to the west, and the weather changes from rainy to clearing, or showery before finally clearing with approach to the anticyclonic area. The temperature is at first quite warm and constant, but with the shifting of the winds to west and northwest, it rapidly becomes colder.

At the north side of the cyclone, proceeding from the front (east) to the rear (west), the wind changes from east through north to northwest. With the east and northeast winds there are clouds and rain, but clear weather with north and northwest winds. The temperature falls rapidly with the beginning of the north wind.

For the anticyclonic area there is usually clear, cool weather, with north and northwest winds on the front side,

and east winds on the rear. On the south (passing from east to west), the winds shift from northeast through east to southeast, the temperature at first falling, and then rising; while on the north they shift from northwest through west to southern, the temperature not varying much. On the west side (passing from north to south), the winds shift from southeast to northeast through the east, the temperature grows warmer, and the clouds of the following cyclone begin to make their appearance.

Methods of Making Weather Predictions. — Weather predictions are made, first, from local observations, and refer to the place at which they are made; second, from weather charts of an extended region, and refer to any region on the chart; third, from weather charts and local observations, and refer to the region at and around the place at which the local observations are made.

Weather Predictions from Local Observations. — It has been found that the probability of the continuation of existing weather is greater than that for a change of weather; and so we may assume that the present state will continue, unless some phenomenon presents itself which foretells a change in the weather. We have such phenomena in the movement and condition of the atmosphere, as shown by the direction of the wind, and kinds of the clouds. If we take as a standard or normal condition dry weather and a clear or but slightly cloudy sky, then the change to generally wet weather and an overcast sky is heralded by the appearance of the cirrus, cirro-cumulus, cirro-stratus, and stratus clouds which almost invariably precede, or we may say outrun, a cyclonic area.

The next point is to locate with some accuracy the center of the approaching storm. This can be done by placing the back towards the direction of approach

of the surface wind, when the region of low air pressure (cyclonic area) will be on the left and a little in advance, and that of higher pressure (anticyclone) on the right and a little in the rear, of the position occupied by the observer. The latter can then make a guess as to which of the quadrants of the storm area will pass over his locality; and, from a knowledge of the general distribution of the meteorological elements in the ideal cyclone or anticyclone, he can foretell with some degree of accuracy what the coming weather will be. For making local predictions for the morrow from the aspect of the sky, the evening at about sunset is a very favorable time.

Observed Shifting of the Wind. — In the case of a cyclone approaching from the west, if the wind gradually changes from the east around through the north to the northwest, then the cyclone center is passing on the south side of the observer, and he encounters in succession the weather conditions on the northern or left side of the cyclone; but if the wind changes from the east or southeast around through the south to the west and northwest, then the cyclone center is passing to the north of the observer, and he encounters in succession the weather conditions on the southern or right-hand side of the cyclone. In the case of a cyclone approaching from the south, when the wind changes from northeast around through the west the observer is on the left or western side of the cyclone; but when the wind changes around through the east, the observer is on the right or eastern side of the cyclone.

When the observer's position lies in the path of the center of the cyclone, the wind at first remains steady in direction, but almost ceases with the passage of the center, and immediately afterwards blows from the proper quarter

on the following side of the cyclone, without the gradual shifting of direction ordinarily noticed. A rapid and great fall in the barometric pressure indicates the near passage of the center of the cyclone.

Introduction of Weather Charts and Storm Warnings. — About the middle of the eighteenth century it was discovered that great atmospheric disturbances known as storms occurred first in the south and the west of the American colonies, and moved in a northeasterly direction over them; but the portrayal of the construction or mechanism of these atmospheric disturbances was not shown until after the first third of the nineteenth century, when, by means of charts showing the simultaneous distribution of the meteorological elements over a large region of country, the existing conditions were brought out, and the movement of translation over a limited portion of the earth's surface could be followed.

The introduction of the telegraph made it possible to collect meteorological data from a large section of country in time to make it of use in following the weather changes over the whole region at the time the events are actually taking place, and also to transmit storm warnings in advance of the approach of the storm. Systematic work of this kind was begun in a limited way about the middle of the nineteenth century, and during the last quarter of the century most civilized countries have had regularly established services for carrying on a work of this nature.

Use of Weather Maps for Weather Predictions. — Briefly stated, the usual method of using weather maps for predicting the weather is as follows: The charted meteorological observations, and especially observations of the air pressure, show us the location, on the map, of cyclones and anticyclones, and the accompanying weather conditions.

These pass over the country from the west towards the east (or sometimes from the south towards the northeast); and, when we see the rate at which they are moving, it can be calculated about where they will be at any time during the next 24 or 48 hours.

Thus we follow the movement of cyclones and anticyclones and their accompanying weather conditions across the country in much the same manner that we can follow the movements of a railroad train if we know its time and place of starting, and its route and speed. But the cyclones vary so much, in intensity, in the paths which they take, and in the velocity of movement, that their positions and conditions can usually be foretold only day by day. Once having fixed the position of a cyclone or anticyclone with regard to any place, we know the general weather conditions at this place as shown by the distribution of the meteorological elements in cyclones and anticyclones.

We have now to see exactly how weather charts are constructed, and then more particularly how they are used for making weather predictions.

Simultaneous Meteorological Observations for use on weather maps are made at certain hours of the day (at present in the United States at 8 A.M. and 8 P.M., 75th meridian time) by regularly appointed observers at the government meteorological observing stations distributed throughout the country. The results of these observations are at once telegraphed to some central office, where specially appointed officials receive them, and by means of them make charts of the condition of the various meteorological elements, and the weather prevailing over the whole region from which such reports have been received.

There are over a hundred observing stations in the United States, and perhaps a dozen more in Canada, from which complete reports of this nature are received by telegraph at the United States Weather Bureau office at Washington twice a day just after the observations are made. In England such reports are received at London; in France, at Paris; in Germany, at Hamburg and other cities; in Austria, at Vienna; in Russia, at St. Petersburg; etc.

Construction of Weather Maps. — In the construction of weather maps there is used as a basis an outline geographical chart of the whole region from which meteorological data are received by telegraph.

Air Pressure. — The barometric pressures reduced to sea level are written down at the stations to which these belong, and then the isobaric lines are drawn. These full, heavily drawn lines (which are placed on United States weather charts at intervals of one tenth of an inch air pressure) have marked on each in large figures the barometric pressure which they represent, and they show at a glance where the cyclonic and anticyclonic areas are located. The cyclone centers are then marked in large letters *LOW* to signify low barometric pressure. The anticyclone centers are marked *HIGH* to signify high barometric pressure.

These isobars are the most important feature of the weather map; and a person skilled in the study of such maps could make good weather predictions from a chart containing them alone, as he is familiar with the usual distribution of the meteorological elements in and around cyclones and anticyclones.

Temperature. — On this same chart the observed temperatures are written down at the stations to which they belong. Isothermal lines are then drawn at intervals of 5° or 10° F. usually. These isotherms are heavy dotted lines, and each has marked on it in large figures the proper temperature.

These dotted lines show at a glance the regions of cold and warm air. Where they bend towards the equator, we know that the local air is relatively cold; and where they bend towards the pole, the air is relatively warm. On the American maps, regions in which marked changes of temperature have occurred during the past 24 hours are inclosed by a heavy dotted line; predicted cold waves are marked C. W.

Wind. — The wind's direction is indicated at each station by drawing an arrow through the center of the station; the arrow flying with the wind. The velocity of the wind is sometimes written down in miles per hour beside the arrow; or lines may be drawn in the tail of the arrow to indicate feathers, the greater number of lines signifying the greater wind velocity. Wind velocities vary too greatly to easily permit the drawing of lines of equal velocities.

Moisture. — The aspect of the sky, and the amount of moisture as indicated by the degree of cloudiness, are shown on the map by a circle drawn around each station. Thus a ○ is made around the station when the weather is fair and the sky is clear, and the space within the circle is filled up as the degree of cloudiness increases; and when it is raining, the circle is entirely filled up.

In the American maps a bar is drawn through the circle (⦶) when the sky is half covered with cloud. A small white center is still left (◉) when the sky is entirely covered with cloud; and the circle is entirely filled up (●) when it is raining. When it is snowing, several bars are drawn through the open circle (▥). The arrows showing the direction of the wind pass through the centers of these circles. *Such symbols are subject to change.*

Also on the American charts, the regions in which rain has fallen in the past 12 hours are shown by a slight shading on the map. In cases of cyclones with low pressure and correspondingly strong winds and much rain, the previous path of the cyclone over the chart is marked by an arrow-headed line, and on it are placed circles with a cross (-⊕-) to indicate the location of the center of the cyclone at the previous times of observation, which are shown by the attached dates.

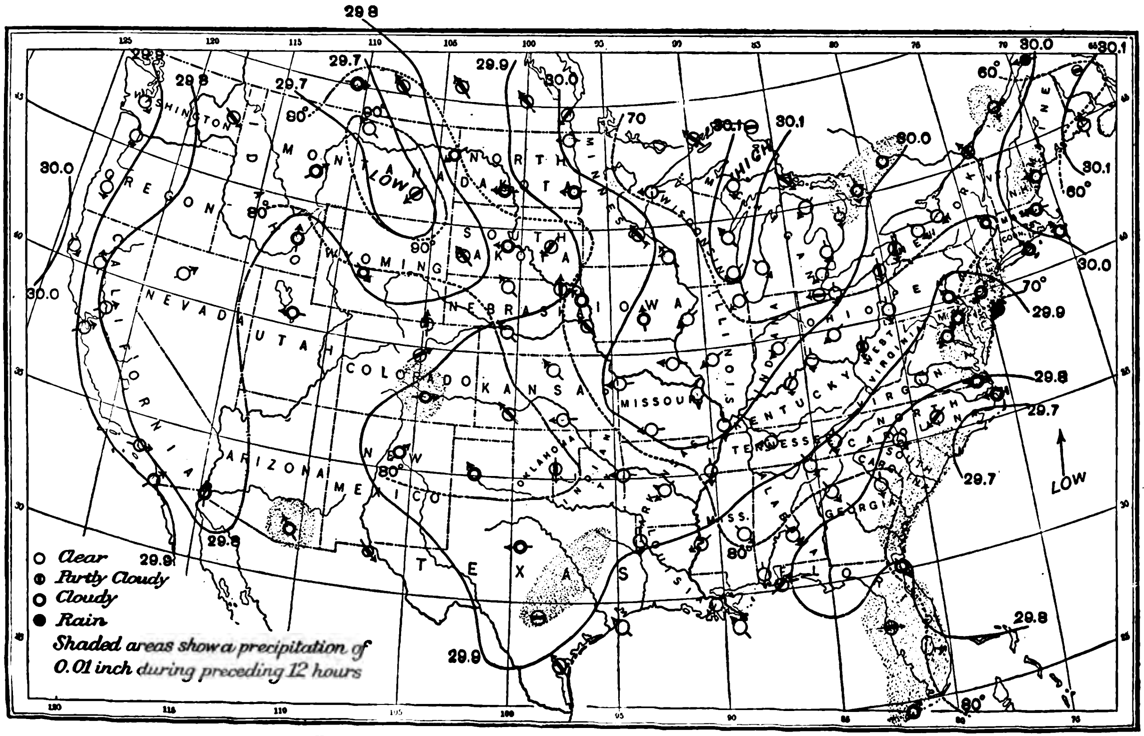


FIG. 87. — UNITED STATES WEATHER MAP, 8 P.M., AUG. 20, 1893.

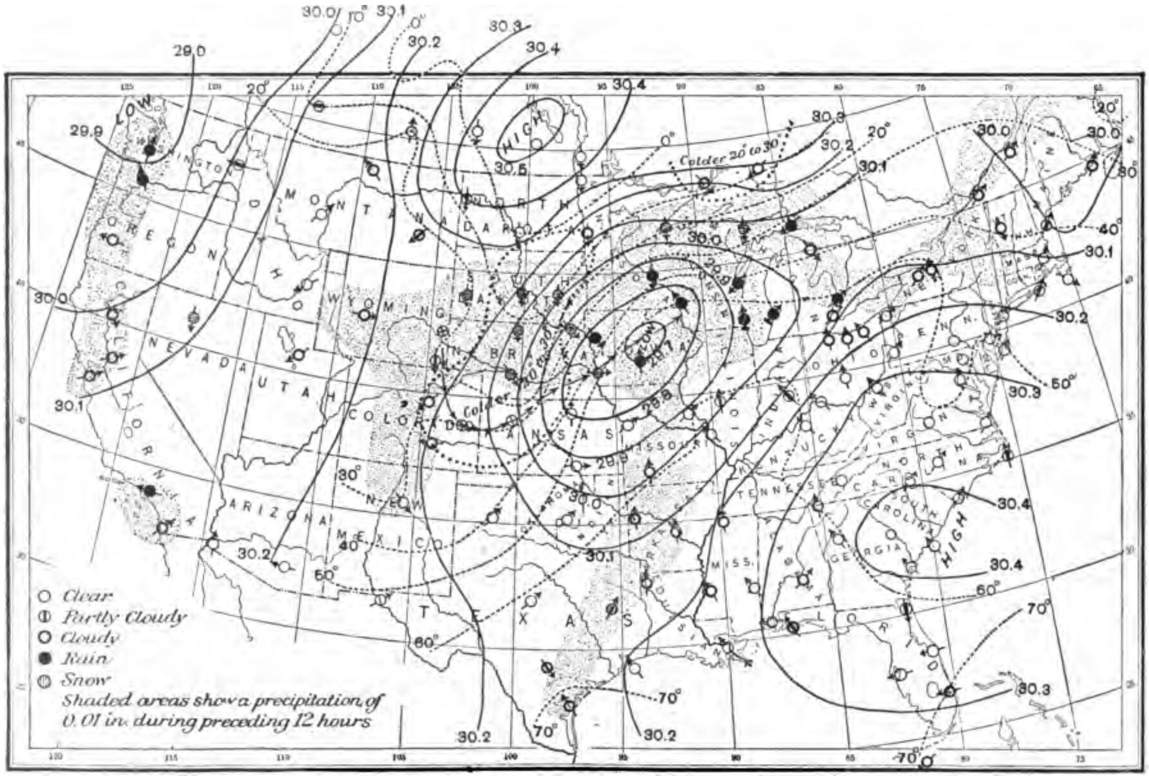


Fig. 88. — United States Weather Map, 8 p.m., Dec. 24, 1893.

Such a weather map shows, first of all, the absolute and relative positions of the areas of high barometric pressure (anticyclones) and low barometric pressure (cyclones), which, in the main, control the characteristics of the weather in our latitudes. It shows the area dominated by the weather conditions pertaining to each. It also shows the magnitude of the excess and deficiency of pressure on which the intensity of the resulting conditions depend. Tabular data accompany most maps.

In order to present the idea of weather charts more clearly, there are given here (Figs. 87 and 88) two actual maps for the United States, showing the simultaneous weather conditions in midsummer and again in midwinter, the two extreme seasons of the year. Very pronounced specimens of the winter and summer types of weather have been selected in order to bring out their characteristics. These maps will be comprehended from the foregoing details concerning their construction, and they should be carefully studied.

Scope of Weather Predictions. — Weather predictions are made either for a single place or for a whole region; and they are made for a definite time, say for 12, 24, or 36 hours in advance, or else the weather changes are designated which are gradually to occur during the coming 12, 24, or 36 hours.

It is best at first to select some definite point for which the weather is to be predicted; and the best one will ordinarily be that at which the person making the prediction lives, because it will then become a matter of practical interest. After some experience has been thus acquired, predictions can be made for some distant point, and still later for a whole region.

Irregular Movements of Cyclones and Anticyclones. — These atmospheric disturbances move (in our latitudes) generally in an easterly direction, and the probability is that the velocity of translation for a few hours preceding any

definite time will remain the same for a few hours after this time; that is, if the cyclone or anticyclone has moved at the rate of 25 miles per hour during the past 12 or 24 hours, then it will probably continue so to move during the next period of like length.

Some weather predictors consider it best to take as the probable future velocity the average velocity of the cyclones, or some compromise between it and the velocity during the hours just passed.

Cyclones entering our continent from the west, and crossing it, frequently move at first in a southeasterly direction, and then in a northeasterly direction; and those which enter the continent from the south or southeast usually move off in a northeasterly direction. Cyclones have a tendency to move slightly spirally around to the right of areas of low temperature, and they have a tendency to drift in the direction of the main atmospheric motion, as shown by the direction of the higher clouds, and to move in the direction of greatest rainfall.

When areas of high barometric pressure (anticyclones) and of low pressure (cyclones) occur simultaneously within a limited region, as on the same weather map, they seem to mutually affect each other's condition, behavior, and movements. When an anticyclone is on the north or east side of a cyclonic area, there is a tendency for the latter to be retarded and deflected towards the north; but if the anticyclone is to the south or southeast of the cyclone, the movement seems to be accelerated. Where one cyclone follows another closely, the one in the rear seems to have a retarding and deflecting effect on the one in front.

After having decided on the probable future path of the cyclones and anticyclones, and their accompanying conditions shown on the weather map, the making of a weather

prediction consists in following out the changes which will take place when the existing conditions are moved along to the places at which they will be if the movement takes place as expected. All of the existing observed conditions may be just moved along over the stationary map in the most frequented path, and at the average rate of movement of the cyclones, and we can see on the map where these conditions will fall.

The conditions actually do change, but it requires great skill on the part of the predictor to foretell these changes.

Variations in Conditions on Weather Charts. — On the first consideration of a weather map, and recognizing that it is made up of areas of high and low air pressure, it might be quite natural to think that in short intervals of time, say every few days, the maps would repeat themselves, and that many would be found quite identical. But this is not the case, for a great variety of combinations exists, and two maps are seldom alike. Still it is possible to class weather maps in a general way according to certain types, but the number of these types is very great. Probably over 100 typical forms would be necessary for such classification; some one of them might occur only at intervals of several years, while others would occur several times in a year.

It has also been found that the behavior of the cyclones and the anticyclones is not the same on maps which appear to be almost exactly alike; so that, because they follow certain movements in one case, it is by no means certain or even probable that the same will occur for other cases of the same type of form and arrangement. It is this variability under apparently like circumstances which precludes a much further advance in accuracy of weather predictions by the methods now adopted.

Weather Predictions for Different Regions. — It must be

assumed, in the case of general weather predictions which are made for a whole continent from a single point (as at Washington), that there will not be much change in the distribution of the weather elements in the cyclonic and anticyclonic areas as they traverse the continent; but they are as a fact continually undergoing changes, which can seldom be anticipated at the proper time. As the cyclones and anticyclones move over the land, new air is brought into the circulation; and it makes a great difference in the future conditions within the areas whether the air thus drawn in has a constant or varying supply of moisture.

Any portion of the cyclonic area, as for instance the southeast quadrant, may have certain definite conditions when the southwest wind which supplies it with air comes from a dry region (as, for instance, in Colorado); but let the area progress eastward, and the air be drawn from a moister region (say, the Ohio River valley), and the weather would not be of such a nature that a prediction for the dry region (Colorado) would fit the existing conditions in the new locality. Let the storm progress still farther, so that the southeast quadrant would be on a warm moist coast (say, on the southern coast of New England), then the warm moist air entering it (from the Gulf Stream region of the Atlantic Ocean) would cause still other conditions to arise.

The meteorological and topographical characteristics of different regions must be studied very closely by the person making predictions for them.

Combination of Weather Map and Local Meteorological Conditions. — It is now a recognized fact that a person on the field of any region can, with the aid of weather maps, make more accurate weather predictions for that region than a person at a distance, who is unacquainted with the local characteristic weather, and relies entirely on the weather map. In addition to the general conditions of environment, there are many local peculiarities which cannot

be taken into account in telegraphic reports, and which greatly affect the weather conditions at a place. It seems best, therefore, to have, for each geographical section where the conditions are similar, a local weather predictor who shall receive telegraphic weather reports, and have the use of maps constructed therefrom, and who shall modify the predictions based on this by the local conditions.

This method is now adopted by the United States Weather Bureau; and it seems to be more successful than the older method, where one predictor at Washington had to furnish from the weather map alone the forecasts for all the various regions of the whole country. This combined method has been in use in Europe since the introduction of telegraphic weather services. There the various countries unite in the interchange of the observed data, so that the predictor in each country can have available the whole continental distribution of meteorological data; but the official predictor for each country makes the weather forecasts for his own country alone. The form of weather maps varies in the different lands, although the data used in all are about the same. This makes the predictions in each case (except for Russia) apply to a limited region only; and this is even more necessary in Europe than in America, for in the former the weather predictions are much more difficult to make, on account of both the variability of the paths of the cyclonic areas in high latitudes, and the fact that the atmospheric disturbances come from the ocean to the westward, whence telegraphic communications are not available.

Thunderstorm Predictions. — Thunderstorms and tornadoes belong to that class of local meteorological phenomena that may be predicted for a general region, but it is impossible to predict their formation at any particular place. When, however, they have once formed, a knowledge of their presence can be transmitted by telegraph or telephone to points which lie to the eastward or northeast of them, and which lie in their probable paths. Undoubtedly more harm than good would be done by predictions of

tornadoes for whole regions in which they are likely to occur. In the case of thunderstorms, however, they traverse such a great section of country, that, when it is found that they are active in a certain section of a cyclonic area, their probable occurrence in the same relative position to this area may be predicted in its farther course; but for any region lying in their path nothing more definite can be said than that local (here and there) thunderstorms may be expected.

Prediction of Cold Waves. — By *cold wave* is meant the fall of temperature over a large area from one day to the next; and this change must be more than temporary and local, for the fall of temperature which occurs during a thunderstorm would not be called a cold wave. Cold waves are nearly always connected with cyclonic and anticyclonic disturbances. They almost invariably occur to the west of a cyclonic and to the east of an anticyclonic area. While cyclonic areas usually move easterly or northeasterly, cold waves generally advance from day to day in a southerly or southeasterly direction. The greatest temperature fall during 24 hours is usually quite near to the center of low pressure in the cyclone, and is most likely to occur a little to the south or west of this center. The direction of most rapid change in the temperatures within a cold wave is usually towards the northwest of the center of the cyclone.

A change of 60° F. in the temperature from one day to the next (24 hours) was observed in the United States only twice from 1880 to 1890; and a change of from 50° to 60° F., on but 16 days in 10 years.

Cold-wave Areas, which may in extreme cases cover 1,000,000 square miles, are usually elliptical, with the long axis extending from southwest to northeast, and parallel to

the long axis of the cyclone in front of it. While the length and breadth are sometimes nearly equal, yet the one axis may be eight times as long as the other.

In the United States Weather Bureau, the name *cold wave* is given when the following temperature changes take place in 24 hours (for instance, from 8 A.M. one day to 8 A.M. the next day): in the northwestern States, from Minnesota to Montana, a fall of 20° F., and temperature going as low as 32° F.; in the central region, extending from Colorado to Maine, and as far south as Tennessee, a fall of 18° F., and temperature going down to 34° F.; in the southern United States, a fall of 16° F., and temperature going down to 36° F. The approach of a cold wave is announced by the United States Weather Bureau by the display of a square white flag with a square black center.

Prediction of Hot Waves. — Hot waves may be referred to their three possible causes: —

1. The ground and the lower layers of air may be heated to an excessive degree by the continuous action of the solar rays during successive clear days; and this accumulation of heat becomes greatest when the days are clear and the nights are clouded. These hot waves are of local character, but may be very extensive.

2. The air of vertical currents may have its temperature raised by the *foehn* process. Such conditions may be expected on the leeward side of mountain ranges, where the air on the windward side is moist. These hot waves are usually confined to narrow bands parallel to the mountains. Similar hot waves also occur in anticyclones, where the descending current is fed by air which has moved upwards in a cyclone, and been deprived of a portion of its moisture before it descends again. Such hot waves follow the course of the anticyclones.

3. Warm air may be blown from warmer to colder regions by the continuation of the winds in the proper

direction during an interval of several days. Such heated conditions of the air may be expected, in middle latitudes, just preceding the great cyclones and following the anticyclones; for in both of these cases the air is blown from lower and warmer to higher and colder latitudes sometimes for days in succession. These hot waves are of greatest extent.

Predictions of Hurricanes. — In the hurricanes which come from the West Indies, and skirt the eastern coast of the United States, the storm center sometimes moves along over the ocean, very frequently follows the Gulf Stream, but sometimes penetrates the Gulf of Mexico; or, pursuing a course somewhat inland, it ravages the whole Atlantic seaboard. Such storms require special predictions.

In the incipient stage of the hurricane there is but a small area affected by it, and at this stage no characteristic signs of its approach precede it at any great distance. It is therefore very necessary for the official weather predictor at Washington to receive as early as possible a telegram from the West Indian Islands, announcing the existence of the hurricane; and additional telegrams should be received stating its progressive movement. The storm center moves so slowly that it may be several days after its announcement before it reaches the coast waters of the United States; and very frequently the wave movement imparted to the ocean water by its violence will reach the shores of the United States before the storm itself is noticeable on the weather maps. Often the passage of such a hurricane far out at sea is made manifest by the waves which it causes to lash the coast hundreds of miles from the storm center, while no trace of the latter is visible on the land weather maps.

The successful predictions of the movement of the hurricanes after they have entered the region covered by the weather maps, is a matter of much difficulty, for their exact direction of motion is too uncertain, and their rate of progress is too variable, to foretell the exact location of the storm's center. The warnings which navigators, especially of small coasting vessels, receive concerning these storms is of incalculable value to them.

When the East Indian typhoons approach the Chinese and Japanese coasts, almost unheralded except at telegraphic points, the loss of life and damage to shipping is very great.

Predictions of River Floods. — During a river flood there is a period during which the water rises, another when it is at its greatest height, and a third when it falls. A flood may then be viewed as a progressive wave which has the culminating point of the flood for its crest. The movement of this crest is followed on the map by the predictor as it proceeds down the river valley; and the time of its appearance is predicted for points below by means of its speed as determined by the velocity of the current at the highest point of the flood. Such predictions can be made usually several days in advance for places lying on the lower part of the river course. When a river is fed by several branches, it is necessary to predict the flood conditions of each effective branch, and then combine these predictions in order to determine the flood conditions for the main stream below the point of union of the branches.

Frost Warnings (Cold-wave Frosts). — When the temperature is high and a cold wave makes it fall considerably below the freezing point, then one kind of a frost occurs; and it will be very general and of a wide distribution, and may begin day or night. When the air is dry, and the temperature does not descend below the dew-point temperature, as in many cases of cold waves, little damage is done by the cold, even if the temperature goes below the freezing point of water; but when the dew-point is high, and dew is deposited and then freezes, the damage to plant growth may be very great, even for a temperature at or just a little below the freezing point of water.

Night Frosts. — The more local frosts are those which, under certain conditions, may occur during the usual mini-

mum temperature at night. Briefly stated, it may be said that night frost is to be feared when observations of the temperature and humidity show that the dew-point lies below the freezing point, and it is expected that the temperature will descend below the dew-point.

Storm Signals. — In all civilized countries where weather services are maintained, special signals announcing the approach of storms are displayed at seaports for the benefit of mariners. These signals are usually flags by day, and lanterns by night.

Distribution of Weather Predictions (especially in the United States). — After the official government weather predictor has decided on the probable weather predictions for the next 24 or 36 hours, this forecast must be placed before the public. This is accomplished in a number of ways. Daily weather maps such as have been described are mailed from various centers to those places which can be reached in time to be of any use, and these are displayed in public buildings, such as post offices.

The telegraph and telephone services are more and more used for distributing weather predictions to distant and local points, and especially in transmitting them to all of the great daily newspapers of the country. It is through these daily papers that the widest-spread knowledge of the weather forecasts is obtained.

Use is also made of the frequent railroad trains in some regions to display on them flags which shall communicate to the inhabitants near the railroad the coming weather conditions; and the railroad telegraph conveys the forecasts to each station on the route, where they are posted up for public information. The press telegraphic organizations transmit the predictions to the city newspapers.

Accuracy of Weather Predictions. — The accuracy of weather predictions is measured in percentage of their

complete success. It is estimated that, on the average, from 80% to 85% of weather predictions are successful. The accuracy varies considerably, not only for the different meteorological elements, but also for different regions of a country. Those regions are the most difficult to predict for which have the greatest and the most sudden weather changes. It is easier to predict the weather during the season of settled weather or least rain than when rainfall is most frequent. It is also easier to predict temperature changes within certain fixed limits in the summer than in the winter.

The predictions of wind velocities and cold waves are about the least successful, averaging only about 60%; while the wind direction is the most successful, averaging over 90%. The weather predictions for the California coast are the most successful of any in the United States.

Long-range Weather Predictions. — While it is possible in our latitudes to make predictions concerning the weather for the next 24 or even 36 hours, and in special cases even longer, yet, in general, the making of definite forecasts for successive days, or for any specified date several days ahead, cannot be successfully accomplished. All long-range predictions, such as for a coming season, are groundless; and the prediction of any great storms for certain dates far in the future is utter nonsense, as no one can foretell the appearance of a particular storm until it actually begins to form. It is possible, however, to state the average weather conditions which have been found to exist at different periods during the year.

CHAPTER XII.

CLIMATE.

Climatic Conditions. — The term *climate* signifies the aggregate or average of meteorological conditions. The climate of a place can be determined only by a series of observations of the meteorological elements, carried on through a period long enough to give the average conditions, freed from the irregularities due to accidental weather conditions, and to determine the average and possible extreme departures from those average conditions.

Weather is but a phase of climate extending over a not too long definite interval of time.

We speak of the meteorological conditions for a single year, or a season, or a day, as weather. Thus we say, "last summer the weather was wet," and not, "the climate was wet." We could not say that the climate was wet during the summer, unless it was so during the majority of summers. We should speak of the annual or seasonal climate, and perhaps also for the months, but not for shorter periods.

Climatology is the science of climate, and its object is to present the average joint action (and possible variations therefrom) of the meteorological elements which pertain to climate at the different places on the earth's surface. Climatology is therefore mostly descriptive. The important climatological elements are the heat, movement, and moisture of the air; precipitation; evaporation; cloudiness; and solar radiation.

The barometric pressure is of the greatest importance in influencing the distribution of the climatological elements; but it is only in the limited regions of great altitudes that it can be considered one of them, and then chiefly on account of its effects on animal organisms, due to the decrease in air density.

In studying climate, we must, then, study the conditions and relations arising from the combination of the average conditions of the above-mentioned elements, which have been studied separately in the earlier part of this book.

The relative importance of the climatological elements, and the data usually required for each, are as follows: —

Temperature.

1. The monthly and annual mean air temperature.
2. The amount of the daily temperature oscillation in the single months.
3. The average and absolute monthly and annual temperature extremes.
4. The average and extreme dates when ice forms in the spring and fall, and the number of days free from ice.
5. The average variability of the daily temperature for each month and the year.

It is desirable, also, to have the average monthly and annual temperatures at some morning, some midday, and some evening hour, say 7 A.M., 2 P.M., and 9 P.M.

Moisture and Precipitation (for each month and the year).

1. The average relative humidity.
2. The average amount of precipitation.
3. The average number of days with precipitation.
4. The average intensity of rainfall.
5. The average probability of rainfall.

Cloud (for each month and the year).

1. The average degree of cloudiness.
2. The average number of hours with sunshine.

Wind (for each month and the year).

1. The average wind velocity or force.

2. The average number of times the wind blows from each of the eight points of the compass.

Evaporation. — It is desirable to obtain the average monthly amount of water evaporated both from a free water surface exposed to the sun and all weathers, and from a shaded water surface; but these are very seldom measured.

Solar Climate. — The ideal solar climate is that which would exist if the sun's rays reached a homogeneous earth without an atmosphere; that is, an earth with a smooth surface all land or all water. The solar climate would have a distribution varying with the latitude from the equator to the pole, but would be the same at all points on a given parallel of latitude. There would exist climatic zones extending around the earth, and following the parallels of latitude.

The Telluric or Physical Climate of the earth is the solar climate modified by the atmospheric conditions, and the existing distribution of land and water surface of the earth. The most important disturbing causes are the unequal distribution of land and water surfaces, and the different elevations of the land surface above the sea level. These give us the three chief forms of climate of telluric origin: 1. The land or continental climate. 2. The water or oceanic climate. 3. The mountain climate.

The fact that the parallels of latitude run partly over a land and partly over a water surface causes differences of climate in an east-westerly direction, transverse to those of the solar climate. These east-westerly differences of climate are reënforced by the transfer of equatorial heat towards the pole, and of polar cold towards the equator, by the currents of air and water.

The Land or Continental Climate is characterized by great extremes of temperature, — relatively high temperatures during the summer (or during the daytime), and low temperatures during the winter (or during the night). The humidity, cloud, and rainfall are deficient.

The Water or Oceanic Climate is characterized by slight temperature extremes: the temperatures remain relatively low during the summer (or during the daytime), and relatively high during the winter (or during the night). The humidity and cloud are excessive, and rainfall ample.

Continental and Oceanic Climates are best considered together, in order to compare the two extreme conditions.

The Temperature. — When the same amount of heat falls on land and water surfaces, the temperature of the land is raised nearly twice as many degrees as the water. The land, then, in summer and in the daytime, warms the air above it more quickly than does the water surface, but the latter gives up to the air above it, by evaporation, more moisture than does the land surface; and when this moisture is condensed in the higher air layers, it gives out heat to them.

At night and in winter, however, the land cools more quickly than the water, and so the air temperatures over the water do not fall so low as those over the land. The greater moisture and cloudiness of the air over the water also prevent the loss of heat as rapidly as over the land. The temperature differences over a land surface in summer and winter are intensified by the decrease of cloudiness towards the interior of a continent, and these differences over a water surface are lessened by the greater cloudiness over the oceans. In higher latitudes a less degree of cloudiness lowers the winter temperatures and raises the summer temperatures. In lower latitudes a decrease of cloudiness increases the temperature.

At about latitude 40°, both north and south, the land and water surfaces have nearly equal temperatures. On the equatorial side of this parallel the land surface, and on the polar side the water surface, is the warmer, on the average, for the year.

At the equator the temperature for a land surface would be about 113° F., and for a water surface 72° F.; at latitude 45°, both land and water would have a temperature of about 50° F.; while at the pole the temperatures would be, for land −25° F., and for water 12° F.

The average temperature of a parallel can be considered as dependent on the latitude and on the relative amounts of land and water.

The Wind. — The differences in temperature between the land and water surfaces give rise to the daily land and sea breezes on the coast, and to the seasonal monsoon winds extending from the interior of the continents far out over the oceans. The winds belonging to the cyclonic and anticyclonic circulation are not influenced very much by the land and sea surfaces, except that the winds become stronger over the water surface because the friction is less than over the land.

The most important climatic effects of the wind are its transference of moisture from the ocean to the land surface; of heated air to colder regions, and cold air to warmer regions; and of air from higher to lower, and lower to higher altitudes, thereby causing adiabatic heating and cooling.

The flow of air between the continents and oceans is much more general and also more rapid in winter than in summer, because the continents are cooled more in winter than they are warmed in summer. The cold winds blow from the continents in winter, and towards them in summer.

The windward coasts of the continent are those towards which the wind blows from the ocean, and they have an oceanic climate. The leeward coasts of the continent are those from which the wind blows towards the ocean, and they have a continental climate. The region to the leeward of isolated large bodies of water, like the American Great Lakes, is warmer in winter and cooler in summer than the region to the windward; because the water surface is cooler in summer and warmer in winter than the land, and the temperature of the air which approaches from the windward of a lake becomes tempered by the water before it reaches the land on the leeward side of the lake.

The Moisture. — The moisture is transferred by the winds from the oceans to the continents. The windward coasts or sides of the continents are therefore moister than the leeward coasts; because on the windward side moist air is received from the oceans, while on the leeward side the air comes from the interior of the continent, where moisture is taken from the air more rapidly than it is added.

The flatter and lower the surface of the land, the farther inland does the air penetrate towards the interior of a continent before it loses its excessive moisture. Where mountain ranges or other elevated lands lie in the path of the moisture-laden air, much of the moisture condenses on the windward side, and causes a wet climate, while on the leeward side the climate is dry. This is most marked when the mountains lie close to the ocean, so that precipitation occurs before a large portion of the moisture has gradually been lost while passing over a large extent of land.

Climatic Effects of the Oceanic Circulation. — The great currents which exist in the oceans, by which the warm water of the equatorial regions is carried towards the pole, and the colder water is transferred from the polar regions

towards the equator, very greatly modify the climatic conditions over the oceans and along coast lands. The warm current heats the air above it, and increases its capacity for moisture; and the cold current cools the air above it, and decreases its capacity for moisture; and the air thus heated or cooled is blown by the prevailing winds on to the continents, or distributed over the adjacent ocean surface. The temperature of the water is communicated to the land, only when the wind blows from the water towards the land.

Effect of Vegetation, and especially Forests, on Climate. Local Effects. *Temperature of the Soil.* — The general influence of the forest is, on the whole, to cool the soil. The extremes of temperature are reduced as compared with those of the open fields, but the effect is more marked (is greater) on the summer maximum than on the winter minimum temperatures.

Temperature of the Air. — The air temperature *under* the crowns of forest trees is, on the whole, cooler — lower in summer, and higher in winter — than in the open fields.

The kind of trees composing a forest has an influence on the temperature, although the average for the year is about the same for all kinds of trees. For evergreen trees, the difference in the temperatures under the trees and in the open shows a symmetrical increase and decrease during the year, being least in winter, and greatest in summer; but for the deciduous trees the difference is variable, diminishing from midwinter to springtime, but increasing rapidly with the growth of leaves. In young forests the maximum air temperatures are lower, and the minimum higher, than in old forests.

The air temperature *within* the crowns of trees is higher than for the same elevation above ground in the open fields.

The difference in temperatures within the crowns and below is more constant in evergreen than in deciduous trees. There is frequently in the woods, especially in summer time, a higher temperature above than below.

The air temperatures above the crowns of trees are similar to those over a grassy meadow or cornfield, — warmer than the air in the sunshine by day, and cooler by night. The effect is greatest for evergreen trees and for deciduous trees in full leaf.

The temperatures within the tree trunk are a little higher than those of the surrounding air in the early spring and late summer, but for the rest of the year are lower.

Local Moisture. — The absolute humidity within a forest slightly exceeds that of the open; and the relative humidity is from 2 % to 4 % greater, the excess being much more marked for evergreen than for deciduous trees. It is not known that precipitation is greater over forests than over the open, but the presence of trees retards the rapidity of loss of fallen rain.

In middle latitudes the annual evaporation from a surface within a forest is about half that in an open field, and the maximum evaporation in the forest occurs in May or June.

Influence of Forests on the Surrounding Climate. — The water which the roots of trees collect from the ground, and which passes through the trunk and branches, is said to be *transpired.* The water which is evaporated from the tree crowns through transpiration is carried as vapor by the winds to the adjacent regions, and the moisture on the leeward side of forests is thus slightly increased.

Local air currents arise due to the difference between the temperatures of the forest and of the surrounding country. Cooler currents in the lower air strata come from

the forest by day, and warmer ones in the upper strata by night.

The forests act as a wind break for the regions close by on the leeward side, and they also reduce the force of the wind in the lower air layers by the friction with the tree tops. This last would consequently reduce the evaporation, which depends so much on the wind velocity.

When the forest forms a glade around a small open area of land, the climate of the inclosed space is not so extreme as it would otherwise be.

In dry regions, when the water is stored underground, and prevented from coming to the surface by overlying hard strata of earth, then the deep penetrating roots of forest trees bring this moisture to the surface, and by transpiration give it to the air.

Altitude and Climate. — The climatic factors of a place — due to its location as regards latitude, and continental or oceanic exposure — are greatly modified by the altitude of the place above sea level. With increase of altitude, the air pressure, the temperature, and the absolute amount of moisture in the air, and above a certain altitude the rainfall, decrease; while the intensity of the direct solar rays, the evaporation, the winds, and up to a certain altitude the rainfall, increase. Since these variations are mentioned under the different meteorological elements in treating them separately, they are not dwelt on further here.

Climatic Zones. — It has been customary to divide the climates of the earth's surface, according to the solar climate, into three zones. The tropical or warm zone lies between the tropics; the temperate zone lies between the tropic and the polar circle, and extends from latitude $23\frac{1}{2}°$ to latitude $66\frac{1}{2}°$; and the polar or frigid zone lies within the polar circle, and extends from latitude $66\frac{1}{2}°$ to the

pole. The temperate and polar zones occur in both the northern and southern hemispheres, and the torrid zone extends from the equator $23\frac{1}{2}°$ into each hemisphere; the areas included in these zones have in each hemisphere the following ratio: torrid zone 5, temperate zone 6.5, polar zone 1.

Tropical Climate. — The characteristic of the climate of the tropical zone is a uniformity of the climatic elements such as exists in no other zone. The periodic changes depending on the daily and annual course of the sun are the most pronounced and regular, while the unperiodic or irregular changes are of but secondary importance. The temperature is remarkably constant throughout the year. The rainfall is copious. There is a rainy and slightly cooler season, and a dry and slightly warmer season, and these are governed by the character of the wind. The rainy season begins about the time when the sun reaches its greatest altitude. The winds are mainly those belonging to the general atmospheric circulation and the monsoons, and are consequently permanent throughout a season. The weather has great permanency of conditions, as the extensive cyclones which cause the frequent and great weather changes in higher latitudes seldom occur in the tropics. During the wet season or during the dry season, the weather of one day is very much like that of another.

The relative and the absolute humidity are great, and the heat is consequently oppressive. During the dry season the sky is usually clear, but during the wet season generally cloudy. During the wet season, after noon, thunderstorms of great intensity occur with great regularity.

The twilight is short, and does not vary much in length throughout the year. Vegetation grows the year round.

Temperate Climate. — The following are the characteris-

tics of the climate of the temperate zones: The temperature is subject to great and sudden changes from day to day; and there is also a great difference in the temperature in winter and in summer. During the winter the temperature is as low as, and in some cases lower than, that of some regions of the polar zone; while during the summer the temperature is in some cases higher than that in the tropical zone. The prevailing winds are from the west towards the east, but the frequent occurrence of extensive cyclonic and anticyclonic disturbances causes a great variability of the wind, both in direction and velocity.

The times and amounts of rainfall are very irregular, as are also the amounts of moisture and cloud. These all depend very much on the local conditions and the temporary character of the wind.

Vegetation grows during at least six months of the year, but during the cold season is at a standstill.

The climate of the temperate zone in the southern hemisphere is much more equable than in the northern hemisphere, on account of the greater amount of water surface in the former. In the temperate zone mankind has reached the highest development.

Polar Climate. — The principal characteristic of the climate of the polar or frigid zone is the absence of solar rays during a longer or shorter portion of the year. The daily temperature change is small. During the winter constant relatively low temperatures prevail, and during the summer constant relatively high temperatures are experienced. The average temperature is lower than that of the temperate zone, but the actual minima are not necessarily lower. The winds are irregular in character, but there is greater frequency of calms than in the temperate zone. The precipitation is slight, and occurs mostly

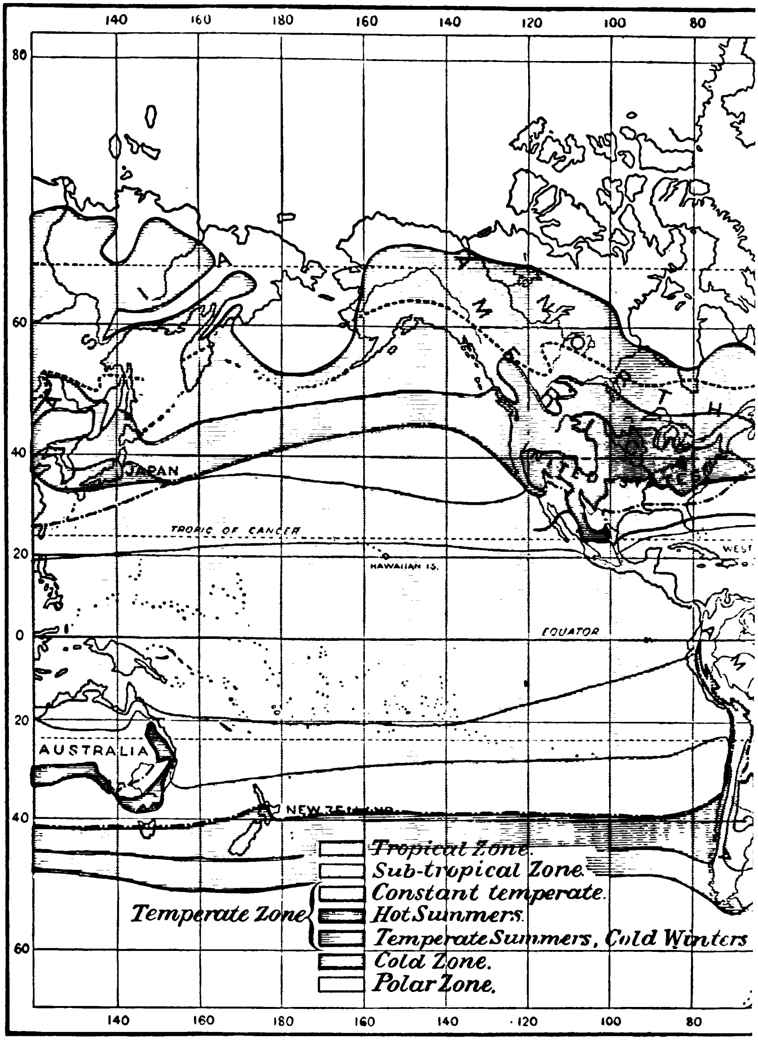
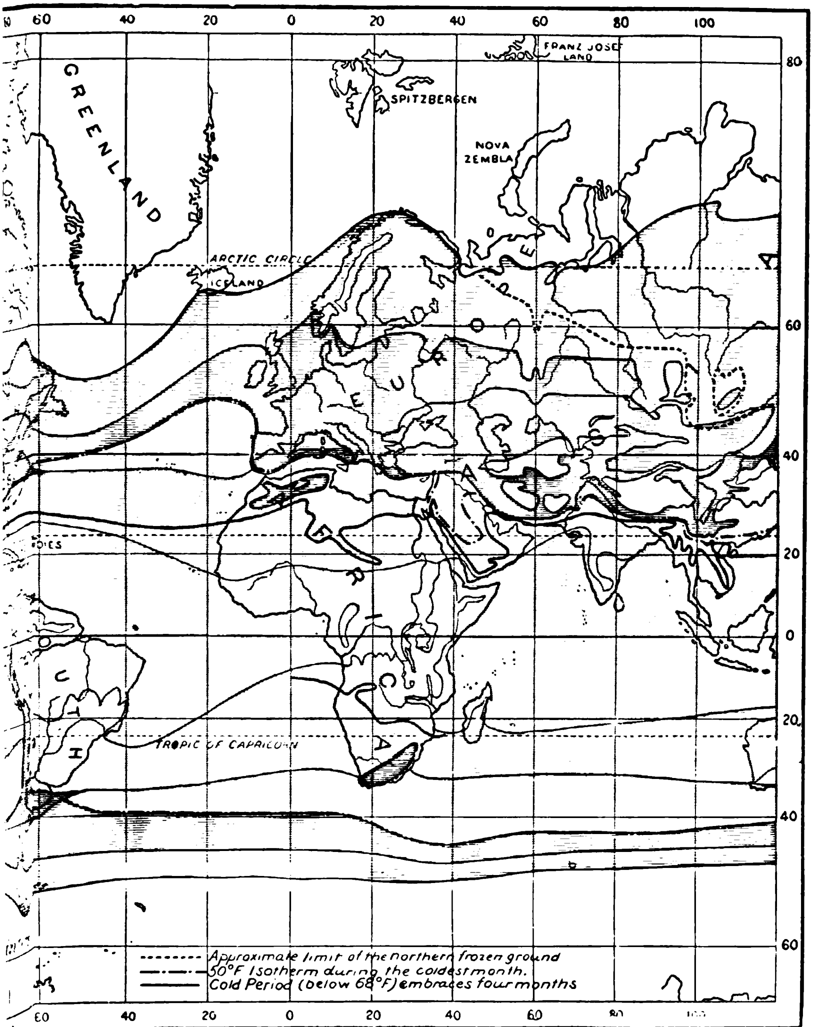


FIG. 89. — HEAT ZONES OF THE EARTH, ACCORDING TO THE DURA



ON OF THE HOT, TEMPERATE, AND COLD PERIODS (AFTER KÖPPEN). (305)

as snow. The absolute humidity is slight, but the relative humidity and cloudiness are sometimes excessive. Only special forms of vegetation grow at all, and these during but a short season. It is the least desirable climate for the development of mankind.

The Heat Zones of the Earth. — It has been found convenient to make a more special division of the earth's surface into seven heat zones, according to the duration of the hot, temperate, and cold periods (Fig. 89, after Köppen). These purely arbitrary divisions are as follows: —

1. *The tropical zone*, which has an average temperature of over 68° F. during all months.

2. *The subtropical zone*, which has during from 4 to 11 months only, an average temperature of over 68° F.

3, 4, 5. *The temperate zone*, which has from 4 to 12 months when the average temperature lies between 50° and 68° F. It has three subdivisions.

(3) *The constant temperate zone*, which has no month when the average temperature exceeds 68° F., or falls below 50° F.

(4) *The temperate zone with hot summers*, which has but few months in which the average temperature falls below 50° F.

(5) *The temperate zone with moderate summers and cold winters*, which has from 1 to 8 months with average temperature between 50° and 68° F., and from 1 to 8 months with temperatures below 50° F.

6. *The cold zone*, which has from 1 to 4 months with average temperature over 50° F., while the remaining months are cold.

7. *The polar zone*, which has, in general, an average temperature of less than 50° F. during all the months of the year.

Climates of the Continents.

Africa. — Africa is the typical tropical continent; and, since it extends from about latitude 35° south to about the same latitude north, only the northern and southern extremities enter the temperate zones, and the main body is in the tropical zone.

Central or Tropical Africa. — The climate of tropical Africa, which includes the whole of the central part of the continent, is warm and moist, with a dry and a wet season. The western side of the continent is characterized by great rainfall and relatively lower temperature, while the eastern coast is drier and warmer. The rainy season on the west coast ranges from August to November, and on the east coast from November to March or April.

Northern Africa is mostly a desert, and the air is dry and warm. The season of rain is from October to March or April.

Southern Africa has a tendency towards a desert climate (as in the north), except on the eastern coasts, where it is more nearly oceanic in character. The eastern coast is the moister and cooler.

Europe. — The climate of Europe is temperate. The prevailing winds are from the west, and in western Europe come from the ocean; and so the western parts, and especially western coasts, are moist and warm, but the eastern parts cool and dry.

In western Europe the winters are mild and the summers cool; and the rainfall is well distributed through the year, but in most regions somewhat more rain falls in the summer and fall than at other times of the year.

In central and eastern Europe the winters are cold and the summers warm. The rainfall is fairly well distributed

through the year, but is greatest in summer. The winters grow colder and the summers warmer, and consequently there are greater extremes of temperature, with the eastward progress through Europe; and the climate changes from oceanic on the western side to continental on the eastern side.

Southern Europe, which borders on the Mediterranean Sea, has a peculiarly warm and rather dry climate, due to the warm winds from the African desert and Mediterranean Sea on the south, and the protecting influence of the Alps Mountains, which keep out the cold winds from the north. The season of greatest rainfall is well marked, and extends from October to February.

Asia. — The climate of Asia is tropical at the south, temperate at the middle latitudes, and polar in the northern portion.

Tropical Asia is that portion south of the Himalaya Mountains, and the extreme southeastern part of the continent. It has a summer monsoon wind blowing from the southwest, and bringing the moist sea air over the land, and producing much rain, especially in India. In winter the monsoon and trade winds blow from the north and northeast, and so from the interior of the continent over India, and render the air there dry and cool; while over the extreme southeastern part of Asia the northeast trade winds bring neither such dry nor such cold air as for India proper. The seasons of change of the wind from one monsoon to the other are characterized by variable winds, hurricanes, and stormy weather. The coldest weather is in December or January, and the warmest from April to July. The daily oscillation of temperature is excessive in the dry season, and is small in the rainy season; but the annual oscillation is not great. The hu-

midity is great in summer, but slight in winter. There are three distinct seasons, — the cool season, from October to February or March; the hot season, from March to July; and the rainy season, from July to October. The vegetation is active throughout the whole year, and is luxuriant.

At the time of change of monsoons in May and October, cyclones of great energy sweep over the Bay of Bengal, and do great damage at sea and on the coast of India.

Temperate Asia, which extends from the Himalaya Mountains and the Yangtse Kiang northward to Siberia, is characterized by a continental climate. The winters are cold and dry, and the summers hot but not very rainy (except on the southern part of the eastern coast, where the rainfall is copious). The annual rainfall of the main region north of the Himalayas is slight. Thunderstorms are infrequent. The humidity is low, and the degree of cloudiness is small.

The summers are warm enough to produce a good vegetable growth where there is sufficient rainfall.

The prevailing winter winds are from the southwest, but in the summer time northeast winds are as frequent as the southwest winds. There are frequent calms in winter.

On the southern east coast the precipitation is great, the climate mild, and warm southwest winds blow. In the winter the monsoon winds are from the northwest. From August to October is the season for the terrible typhoons of this coast.

Polar Asia. — The polar climate of Asia extends from the Arctic Ocean practically to nearly latitude 50°. It extends so much beyond the Arctic Circle, on account of the marked continental features of the climate. The winters are long and intensely cold; the summers short and

warm, except in the extreme north, where the summer heat is not so great. The periodic daily temperature oscillation is large.

The absolute and the relative humidity are slight, and the degree of cloudiness small for an arctic climate; the winter sky being clear. The precipitation is slight. In the winter the winds are from the northwest, and in summer from the south and southeast. The short, hot summer is favorable for the hardy vegetable growth, and trees and grains grow up to a very high latitude.

On the extreme coast the winters are milder, the summers cooler, and the moisture, cloudiness, and rainfall increase.

South America. — The South American continent extends from 10° north latitude to about 50° south latitude. It has therefore a tropical climate, except for the southern portion, which extends far into the temperate zone.

Tropical South America and *Central America* have a warm, moist climate. The range of temperature during the year is slight except just south of the United States, where the cold north winds from central North America make themselves felt to within 15° of the equator. The rainfall is excessive except on the west coasts, where the high Cordilleras prevent the moisture with which the prevailing east winds are laden from reaching the coast. There is, then, a strip along most of the west coast where dry conditions prevail. The greatest rainfall occurs in general in Brazil in the summer season of the southern hemisphere, and during the months of May to August little rain falls; but the varieties of rainfall types throughout this whole tropical region are numerous.

Temperate South America. — The temperature on the east coast is higher than on the west coast, and inland it

is higher still. While the summer temperatures are high, the winter temperatures are not nearly so low as for a similar latitude in North America, because the winds from the ocean temper the climate. On the west coast the annual variation of the temperature is less than in the eastern part of the continent.

The rainfall is greatest on the eastern side in lower latitudes, where the summer is the rainy season, and on the western side in the higher latitudes, where the winter is the rainy season. At the southern point of the continent the rainfall occurs about equally at all seasons. In lower latitudes the winds on the east side are variable, but are mostly from the north and northeast; but on the west side south and southwest winds are most frequent. In higher latitudes the prevailing winds are from the west.

On the southern extremity of the continent the humidity and the cloudiness are great, and the temperature equable, owing to the oceanic character of the climate.

North America. — North America has three main types of climate, — the tropical at the south, the temperate occupying the greater portion of the continent, and the polar climate at the north.

Tropical North America is characterized by hot, rainy summers and cool, dry winters. Except where high mountain ranges interfere, the rainfall is copious. The cloudiness and moisture are in excess in the summer, and deficient in winter.

Temperate and Arctic North America. — The climate of the main portion of North America changes gradually from a polar climate at the north to tropical at the south. There are also three climatic zones to be met with in crossing the continent from the east to the west.

The eastern zone extends from the Atlantic coast to the

center of the continent, and has hot summers and cold winters. The humidity is great, and the rainfall is not excessive, nor is it deficient. On the east side of the continent the cold but not very dry polar climate extends southward to low latitudes, and the hot, wet, tropical climate extends far north into relatively high latitudes, so that there are great differences in temperature with slight changes of latitude. Frequent irregular changes of temperature of great magnitude occur.

The central zone extends from the center of the continent to the Pacific coast range of mountains. The winters are cold, the summers hot, the moisture is deficient, and there is little rainfall. The irregular changes of temperature are sudden and great, and the daily amplitude of temperature oscillation is also great.

The western zone embraces a long narrow strip extending from the coast to the nearest high mountain range. At the center and north the winters are not cold, nor are the summers very warm, although intense heat is sometimes experienced; and the humidity and rainfall are excessive in winter, but deficient in summer. In the extreme south the climate is tropical.

The prevailing winds in the eastern half of America are from the northwest in winter, and from the southwest in summer. In the western part they vary greatly for different regions in winter, but are mainly from the west and south in summer.

CHAPTER XIII.

CLIMATE OF THE UNITED STATES.

(This chapter may be considered as an appendix.)

Climatic Location of the United States. — The United States lies almost wholly within the north temperate zone, but extends somewhat into the subtropical zone on the south, and into a cold climate on the north. Since it also extends across the continent, and has oceans on the east and west, and contains extensive mountain ranges, there is a great diversity of climate within its boundaries.

Main Types of Climate in the United States. — There are found the three main continental types of climate properly belonging to middle latitudes, where the prevailing winds are from the west. At the western part, bordering on the Pacific Ocean, there is the relatively mild windward (west) coast climate; at the east there is a leeward (east) coast climate; and at the interior of the continent between these two regions is the severe continental climate. On the southern border the climate becomes almost tropical in character in all three of these longitudinal zones.

The high mountain ranges lying nearly parallel to the Pacific coast and very near to it prevent the extension of the western coast climate to any great distance inland, while the relatively low mountains in the eastern part do not prevent the interior continental climate from extending close to the eastern or Atlantic coast. Thus the greater part of the climate of the United States is that pertaining to the interior of a continent in middle latitudes.

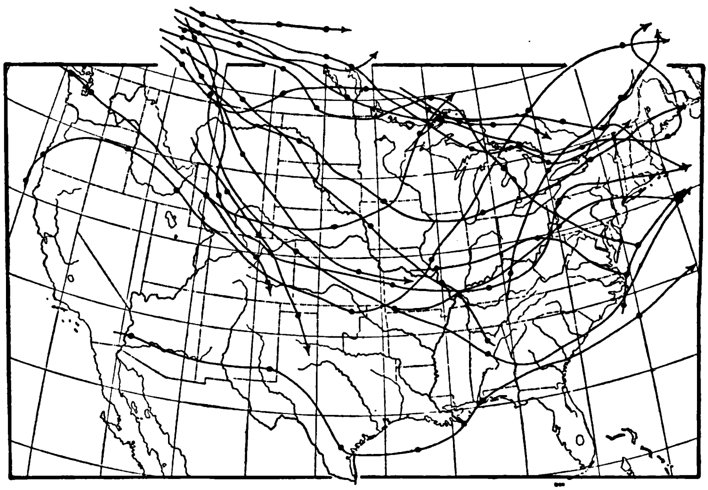


FIG. 90.—PATHS PURSUED BY THE CENTERS OF CYCLONIC AREAS, JANUARY, 1893 (U.S. WEATHER BUREAU).

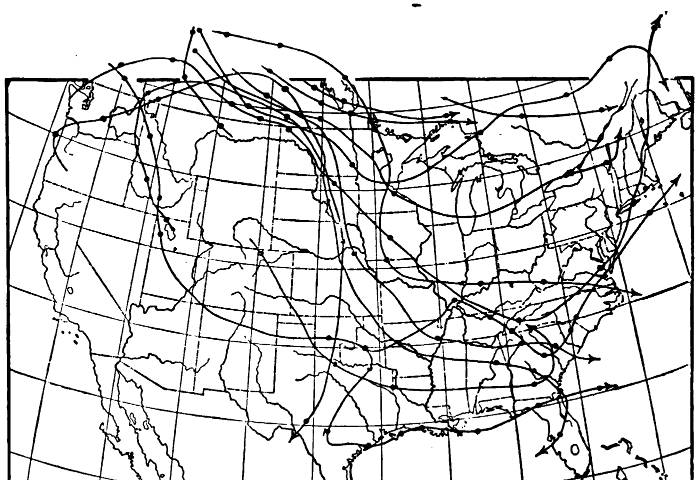


FIG. 91.—PATHS PURSUED BY THE CENTERS OF ANTICYCLONIC AREAS, JANUARY, 1893 (U.S. WEATHER BUREAU).

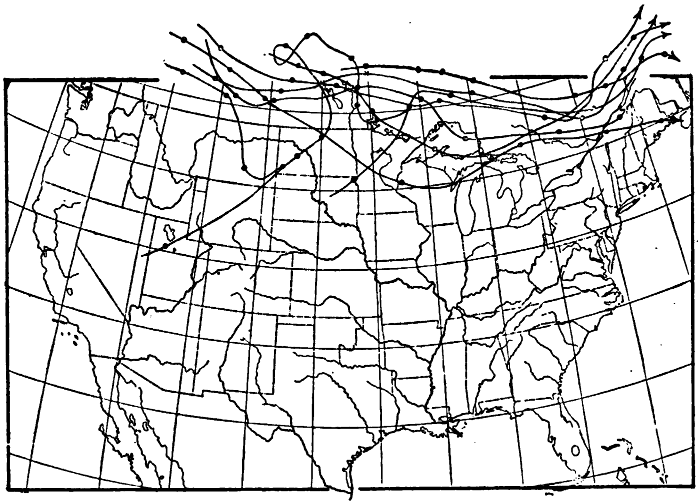


FIG. 92.—PATHS PURSUED BY THE CENTERS OF CYCLONIC AREAS, JULY, 1893 (U.S. WEATHER BUREAU).

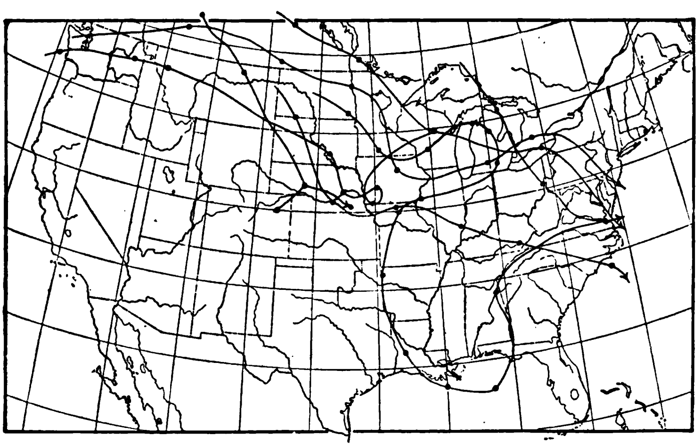


FIG. 93.—PATHS PURSUED BY THE CENTERS OF ANTICYCLONIC AREAS, JULY, 1893 (U.S. WEATHER BUREAU).

Climatic Effects of Cyclones and Anticyclones. — There is another feature which much modifies the normal conditions of climate under these circumstances; and that is the relative frequency, and the location of the paths, of the great cyclonic and anticyclonic disturbances of middle and higher latitudes.

The region of greatest frequency of cyclonic areas lies in the neighborhood of the Great Lakes, and extends from perhaps the 95th meridian through the St. Lawrence valley. To the south and west of this region there is a rapid decrease in their frequency. Since the passage of all cyclonic areas, and the almost invariably accompanying anticyclones, is marked by the succession of meteorological changes already described, the variability of the climatic conditions increases in proportion to the number of cyclones. We thus have the most constant and stable climate in the regions least frequented by the cyclones.

Cyclones are not only much more frequent, but are also much more widely distributed over the country, in the cold season than in the warm. In the latter, however, the path of greatest frequency lies much farther north than in the cold season. And thus, while in the winter the changes due to the passage of the cyclones frequently extend to the Gulf of Mexico, yet in the summer the Southern States do not feel their influence so much, since they are relatively seldom visited by these atmospheric disturbances.

Figs. 90, 91, 92, and 93 show the paths pursued by the centers of cyclonic and anticyclonic areas during the months of January and July, 1893.

The arrows point in the direction of motion, and the distances between the dots on the lines show the movements of translation during 12-hour periods.

Since cyclones and anticyclones control the weather to distances of hundreds of miles from their centers, it is seen that in the cold season almost the whole of the United States comes at times within their limits of domination. In the warm season, although the paths lie farther to the north, yet the southern part of the United States sometimes comes within their influence.

The shifting of the direction of the wind, and the transition from clouded to clear sky, due to the passage of these cyclones and anticyclones, cause the sudden changes in weather which produce the variability of climate so noticeable in the eastern half of the United States. In this latter portion of the country there is sufficient moisture (brought by the winds from the bodies of water on the north, east, and south) to allow the full activity of the cyclones in their most complete forms to make itself felt by producing the characteristics already noticed for cyclones; especially those pertaining to the rapid spread of the rain area and copious rainfall.

In the western and drier half of the United States, however, the lack of moisture and the breaking-up of the winds by the mountain ranges not only prevent the full development of the characteristics of the cyclonic areas, but also render their effects less marked. When the air currents do not supply much moisture, the growth of the rain area is not so rapid, and the rainfall is less abundant.

Climatic Subdivisions of the United States. — The following systematic division of the United States into climatic subdivisions was adopted by the United States Weather Bureau, and is based on the variations of climate with latitude, altitude, and marine exposure (Fig. 94).

Starting with the Pacific coast and passing towards the east, we find six nearly parallel zones of quite distinctive

climatic character, stretching across the country from nearly north to south. These zones are subdivided latitudinally into northern, middle, and southern portions.

1. **The Pacific Coast Region** is divided into the northern, middle, and southern parts, and is the region lying along the Pacific Ocean, and extending perhaps 200 miles inland. It has a very equable temperature due to the prevailing west winds from the ocean. The winters are abnormally warm over the northern and middle sections. The summer heat,

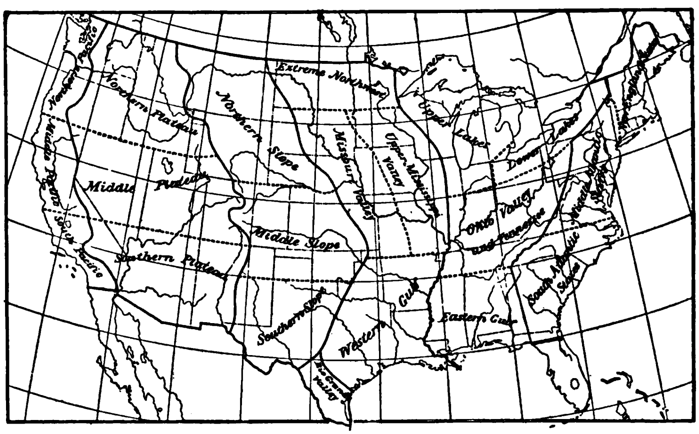


FIG. 94. — THE MAIN CLIMATIC SUBDIVISIONS OF THE UNITED STATES (ADOPTED BY THE U.S. WEATHER BUREAU).

except in the northern portion and on the middle coast, is sometimes excessive. Frosts seldom occur. The year is divided into a wet season in winter and a dry season in summer. The amount of precipitation is very great at the north, and slight at the south, where irrigation is necessary for the cultivation of crops.

2. **The Plateau Region** extends longitudinally from 800 to 1,000 miles between the Sierras and Rocky Mountains,

and is divided latitudinally into the northern, middle, and southern plateau. It has an altitude of several thousand feet, and the air is cool and dry. During the winter the cold is continuous. In summer the high altitude and clear sky allow high temperatures to be reached in the daytime, but at night the radiation is rapid, and the temperature becomes low, so that the daily range of the temperature is great. Hot winds of a *foehn* character occur frequently in the valleys. The moisture and precipitation are deficient throughout the whole region, and agriculture must be carried on mostly by irrigation. The high altitude causes the average cool northern climate to extend far into the southern plateau.

3. **The Eastern Slope of the Rocky Mountains, or the Great Plains,** extends about 500 miles longitudinally from the Rocky Mountains to the low prairie lands of about a thousand feet altitude, and is divided into the northern, middle, and southern slopes. It has a truly continental climate, which is characterized by cold winters and hot summers. Great and sudden changes of temperature occur. In winter, extreme cold occurs on the northern slope, and this cold air is often carried far into the southern slope (Texas) by the north wind at the rear of cyclones. The precipitation is slight at the west, where irrigation is necessary for agriculture, but increases towards the east. The greatest rainfall is in early summer on the northern slope, and in late summer on the southern slope. The average wind velocities over the whole slope are excessive, mainly owing to the lack of obstructions to the wind, such as mountains and forests.

4. **The Central Prairie Lands** extend to the eastward of the eastern slope longitudinally 400 or 500 miles to the Mississippi River, and are divided as follows: the Missouri

valley, the Upper Mississippi valley (the northern part, the Red River and Upper Missouri valleys, is sometimes called the Extreme Northwest, or simply Dakota), and the Western Gulf Region lying to the west of the Lower Mississippi. These have a continental climate with cold winters and hot summers in the Extreme Northwest, the Missouri, and Upper Mississippi river valleys; but in the Western Gulf Region the Gulf of Mexico exerts a modifying influence when the wind is from the south, while the "northers" bring the cold from the north into this region in winter. The rainfall is deficient at the north, but increases to a normal amount (about 40 inches) towards the center, and becomes excessive near the Gulf of Mexico. The winds are strong, and usually from the west in the northern part; but in the Western Gulf Region they are weaker, and move from the east and south. The Rio Grande valley has a climate partaking of some of the characteristics of both the Western Gulf Region and the Southern Slope.

5. **The Western Appalachian Slope** extends to the Mississippi River, and includes the Upper and Lower Lake Region at the north, the Ohio and Tennessee valleys at the center, and the Eastern Gulf Region at the south. In the Lake Region there is a cold winter, but not a very hot summer, the extremes being tempered by the large bodies of water. The winds are strong, and mostly from the west.

In the Ohio and Tennessee river valleys the summers are hot, and the winters short and of but moderate coldness. The winds are not excessive, and blow mostly from the west and southwest.

In the Eastern Gulf States the winters are very short, and ice seldom forms. The summers are very long and hot; but the great amount of forest, and the winds from

the Gulf of Mexico, somewhat temper the heat. The winds are from northern directions in winter, and from southern directions in summer.

The rainfall is about normal, and most frequent in the early summer, in the Lake Regions and Ohio and Tennessee valleys; but it is excessive, and most frequent in late summer and early fall, in the Eastern Gulf States.

6. **The Atlantic Slope** has at the north the New England region, at the center the Middle Atlantic States as far south as North Carolina, and at the south the South Atlantic States, except the middle and southern parts of Florida.

The climate of the Atlantic coast does not differ much from that of the western Appalachian slope at the same latitudes. The winters are quite severe on account of the prevailing west and northwest winds, which blow the cold air from the inland towards the coast. The summers are warm because the winds are from the west and southwest, and these blow the hot continental air coastward. The prevailing winds being towards the ocean (except at the extreme south), the moderating influence of the oceanic air is little felt. The rainfall, which is well distributed through the year, increases from the north towards the south.

Central and southern Florida has an insular, almost tropical climate.

Geographical Distribution of the Climatological Elements over the United States.

This is best shown by means of charts, giving the distribution of the climatic elements. The charts are supplemented by a few remarks concerning the main features of this distribution.

Temperature.

Temperatures of the United States. — Outside of the tropics, the average temperature for the year does not fully represent the climate of a place or region, because the winters are cold, and the summers are warm, and the average annual temperature does not show this oscillation. But the average temperature for the coldest month, and also for the warmest month, does show the annual extremes of seasonal temperature, and the average of the two is very nearly the average temperature for the year.

Since the United States lies mainly in a single solar zone (the north temperate), the coldest and the warmest months are practically the same over the whole country. We shall therefore take the temperatures for the midwinter month of January, and the midsummer month of July, as indicative of the extreme average seasonal temperatures. In addition will be given the maximum and minimum temperatures, the period (number of days) during which the temperature remains below the freezing point of water, the variability or average changes of temperature from one day to the next, and the times of earliest and latest frost.

The January Average Temperatures in the United States are shown on the accompanying chart (Fig. 95). The coldest region, with a temperature of 5° F. below zero, is found near the center of the continent, where the outward radiation is greatest and the warming influence of the ocean is least felt, and lies in North Dakota and Minnesota, and extends up into British America. From this region the temperatures increase towards the south, west, and east. The natural southward increase with decrease of latitude is quite regular, and continues to the Gulf of Mexico, where a temperature of 55° F. is found; but in

a southeast direction it reaches 70° F., at Key West, Fla. To the east of the central cold region, since the prevailing westerly winds carry the interior continental cold almost to the coast, the increase is gradual, to the temperature of 5° F. or 10° F. on the Atlantic coast in the same latitude. To the west, however, the increase is quite rapid, to 40° F. on the Pacific coast. This rapidity of

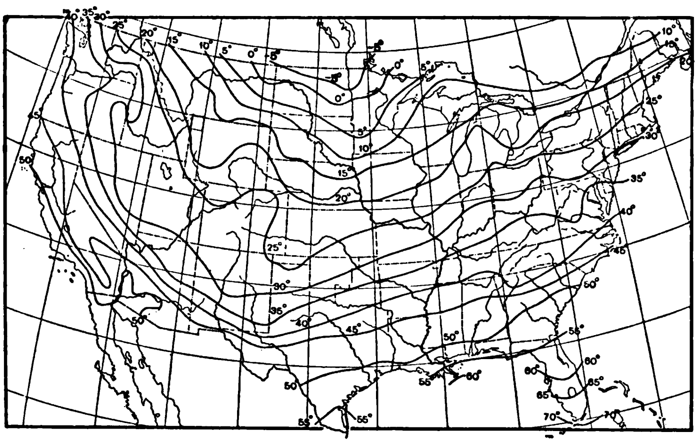


FIG. 95. — AVERAGE TEMPERATURE FOR JANUARY IN THE UNITED STATES (AFTER GREELY).

change is due to the abnormal warming of the western coast by the ocean wind from the west. Towards the southwest the increase is at first rapid, and then slow through Utah and Colorado, owing to the altitude, and then rapid to about 55° F. on the Pacific coast.

Along the Pacific coast there is a decrease of temperature from 55° F. at latitude 33° north, to 40° F. at latitude 48° north; while on the Atlantic coast the temperature at latitude 25° north is 70° F., at 33° north is 50° F., at 45° (latitude of central Maine) is 20° F., and at latitude 48° north is probably about 5° F. This difference between the two coasts is because of the prevailing winds from the west, which give

to the west coast more of the character of an oceanic climate. The Gulf coast region to the west of the Mississippi River is considerably colder than that to the east, the difference reaching even 10° F. at the same latitude in the extreme parts of these regions, because the abnormal cold at the center of the continent influences the coast region very strongly by means of the cold winds from the north.

The coldest month is January, but the coldest days may occur in any other winter month.

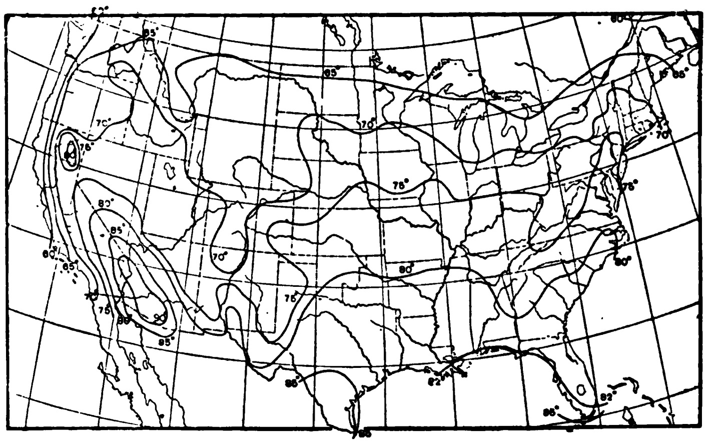


FIG. 96. — AVERAGE TEMPERATURE FOR JULY IN THE UNITED STATES (AFTER GREELY).

The July Average Temperatures (Fig. 96). — In this midsummer month the average temperature is about 65° F. along nearly the whole northern boundary of the United States. There is a gradual increase with southward progress from this northern limit, to about 82° F. on the Atlantic and eastern Gulf coasts, to 85° F. on the western Gulf coast, and to over 90° F. at the base of the southern plateau (in southern Arizona and southeastern California). Along nearly the whole Pacific coast there is a narrow belt having a temperature of but 60° F., due to the winds from the relatively cool Pacific Ocean.

On this chart we see the effects of the abnormal heating of the land surface in summer. The coasts are not now warmer than the interior for the same latitude; and the isotherms bend poleward with removal from the coasts.

The isotherms on the Pacific slope lie farther apart in January than in July, which indicates a much more gradual change of temperature for a given distance in winter than in summer. In the rest of the country the isotherms lie farther apart in July than in January, and the change of temperature is more gradual in summer than in winter. Note the effect of altitude in the Rocky Mountain region in carrying the summer isotherms southward.

The warmest month, but not necessarily that with the hottest days, is July; but there are some exceptions, as, for instance, on the Pacific coast August has the highest average temperature.

Extremes of Temperature. — *The absolute maximum temperatures* observed in the shade in the United States are shown on the chart, Fig. 97. The variations in the extreme maximum temperatures over most of the inland are surprisingly slight. It is seen, however, that in the region which is deficient in moisture, and where the summer sky is almost cloudless, the maximum temperatures rise 5° or 10°, or even in the very dry deserts 20°, above the average maximum temperatures for the whole country, which may be put at a little over 100° F. Along the ocean coasts the modifying influence of the cooler sea reduces the average maximum temperature for the whole country by about 10°; so that the excess caused by desert conditions about equals in magnitude the deficiencies due to the oceanic influence. The climatic influence of the prevailing winds is not markedly felt on the maximum

temperature except very near the coast, and even there the diurnal land and sea breezes play the important rôle. In most of the eastern half of the United States 100° F. is reached except on the New England coast and in the Appalachian Mountains. To the west of about the 95th meridian, temperatures of 105° F. are reached, except on the Middle and Southern Plateau, where they are not over

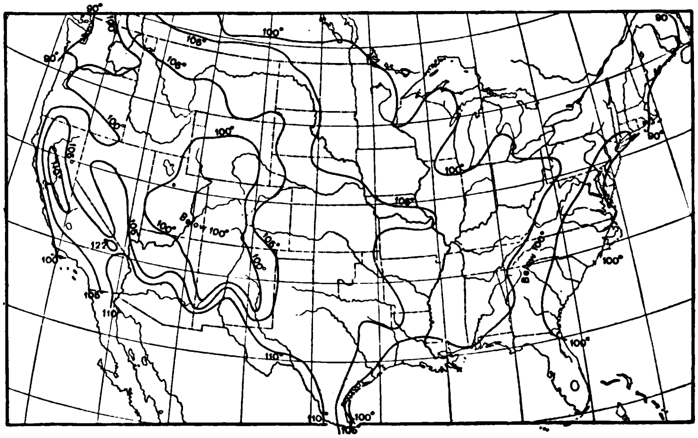


FIG. 97.—ABSOLUTE MAXIMUM SHADE TEMPERATURE IN THE UNITED STATES (AFTER GREELY).

100° F. In central California temperatures of 110° F., and in southeastern California 122° F. (or higher) and about the same in southwestern Arizona, are reached. On the Pacific coast they vary from 90° F. at the north to 100° F. at the south. These high temperatures are much more easily borne in the dry air of the west than are lower temperatures in the moist air of the eastern United States.

The Minimum Temperatures observed in the United States have much more sharply defined limits than the maximum

temperatures just mentioned, and are shown in Fig. 98. The relative general distribution of the minimum temperatures is much the same as that for the January average temperatures, shown on Fig. 95.

At the center of the continent, where the radiation of heat is strongest, there is a region of lowest minimum temperatures, but there is a moderating of these extreme

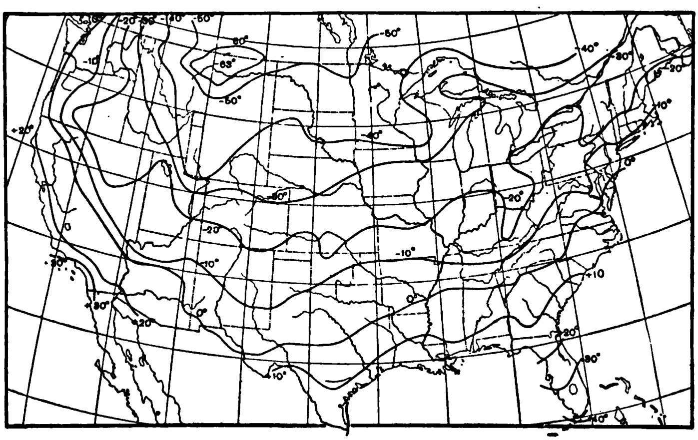


FIG. 98.—ABSOLUTE MINIMUM SHADE TEMPERATURE IN THE UNITED STATES (AFTER GREELY).

temperatures toward the east, west, and south; i.e., toward the coasts and the equator. The general direction of the winds influences this distribution most strongly. Blowing as they do from the west, they make the temperatures on the Pacific coast, where the winds blow from the ocean, from 20° to 30° higher than on the Atlantic coast, where the cold air is blown from the center of the continent eastward. These extreme minimum temperatures in the central and eastern United States occur in the extended cold waves which sweep over the land. The great extent and

regular progressive motion of these cold waves cause the existing regularity in the line of equal minimum temperature in the central and eastern United States.

The extreme minimum of $-63°$ F. observed in northern Montana (Fort Assiniboine) is surrounded by higher temperatures to the east, west, and south. Towards the west from Montana there is at first a gradual and then a more rapid increase to $0°$ F. near the coast, and on the North Pacific coast itself the minimum is from $+8°$ to $+15°$ F. above zero. Towards the east from Montana there is a gradual increase to $-25°$ F. or $-30°$ F. on the Atlantic coast in the same high latitude. Towards the south from Montana there is an increase to $+13°$ F. in southern Texas; towards the southeast there is an increase to $+41°$ F. at the southern end of Florida; and towards the southwest there is an increase to $+32°$ F. on the southwestern coast of California.

On the summit of Mount Washington in New Hampshire the minimum temperature reached is $-50°$ F. at an altitude of about a mile, and on Pike's Peak, Col., about $-40°$ F. at an altitude of about two miles and a half.

The Absolute Oscillation of Temperature (Max.–Min.) is shown in Fig. 99.

This chart brings out very plainly the fact that the amplitude of the temperature depends to a very great extent on the minimum temperatures reached, and is but slightly influenced by the maximum temperatures reached. The amplitudes are greatest at the interior of the continent, but decrease with approach towards the coast and with the latitude; although this latter decrease may be almost masked on a windward coast, even for so great a breadth of latitude as the width of the United States. On the leeward (eastern) coast of the United States the conti-

nental influence is shown by the decrease with latitude coming halfway between that for the interior of the continent and the windward (Pacific) coast.

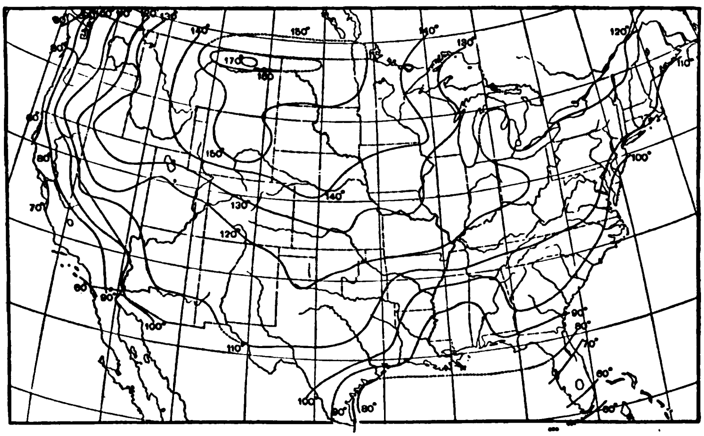


FIG. 99. — ABSOLUTE AMPLITUDE OF OSCILLATION OF SHADE TEMPERATURE IN THE UNITED STATES (AFTER GREELY).

The Duration of Temperatures below Freezing in the United States is shown by the following chart (Fig. 100), on which is given the number of days when the average daily temperature was below 32° F. The most southern line drawn on the map, and marked 0, is the dividing line between the regions where the average daily temperature goes below freezing, and those where it does not. The next line, marked 30, passes through the localities which have 30 days with the average temperature below freezing. The additional lines are similarly drawn for the successive multiples of 30 days (i.e., months).

This distribution follows in a general way that of the midwinter average (Fig. 95) and minimum (Fig. 98) temperatures already described. In northern Minnesota the

number of days is 165, and there is a decrease from there in all directions in the United States. The decrease is very gradual to about 120 days on the northern New England coast on the east, but more rapid to 0 days on the Delaware coast on the southeast, central Texas on the south, central Arizona on the southwest, and central Oregon on the west.

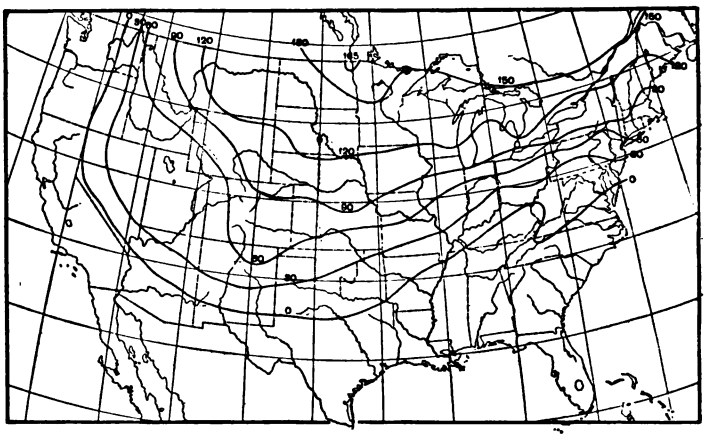


FIG. 100. — NUMBER OF DAYS WITH AVERAGE TEMPERATURE BELOW FREEZING IN THE UNITED STATES (AFTER GREELY).

Variability of Temperature in the United States. — The change in the average temperature from one day to the next varies greatly in the United States, not only during short intervals of time, but also on the average for each month from month to month. Such changes are much greater in some regions than in others.

The variability of the temperature depends largely on the descent of the minimum temperatures, and follows quite closely the distribution of the minimum temperatures (Fig. 98). The variability is greatest in the interior

of the continent, but decreases towards the coasts, and towards the windward (western) coast much more rapidly than towards the leeward (eastern) coast. There is also a decrease towards the equator, scarcely perceptible on the windward coast, very marked at the interior, and between the two on the leeward coast.

In general, the variability is greatest in midwinter (January and February), and least in midsummer (July and August). The frequency and magnitude of these changes are mainly dependent on the number, rapidity of movement, and intensity of the areas of high and low barometric pressure which sweep over the country. The accompanying chart (Fig. 101) shows the average temperature variability for the maximum month of January. It is seen that the greatest variability occurs in the central interior region of the northern part of the United States, where it is about 10.5° F. Towards the west of

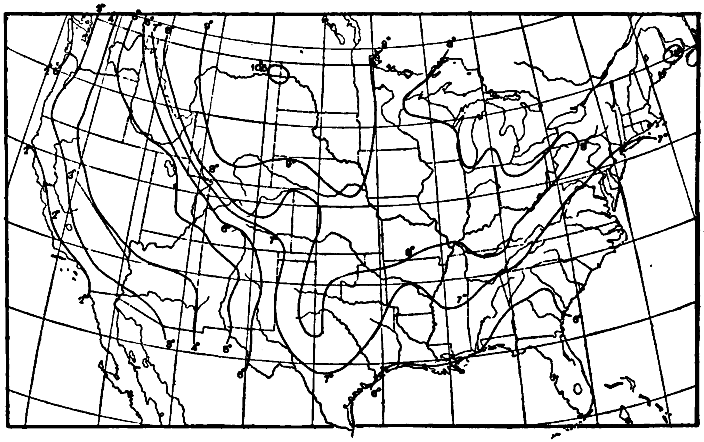


FIG. 101.—VARIABILITY OF AVERAGE DAILY TEMPERATURE IN JANUARY IN THE UNITED STATES (AFTER GREELY).

this region of maximum change, the variability decreases at first slowly, and then more rapidly, to only 2.5° F. on the northern Pacific coast. Towards the southwest the decrease is to 2° F., on the southern Pacific coast. Towards the south the decrease is to 6° F. on the Gulf coast; towards the southeast, to 3° in southern Florida; while towards the east there is but a slight, if any, decrease to the North Atlantic coast.

The increase of the average temperature variability with latitude is scarcely perceptible on the Pacific coast and in the extreme western United States in general; it is about 1° for each 300 miles at the center of the continent; and on the Atlantic coast the increase is quite rapid on the northern and southern coasts, but very gradual on the middle coast.

Relative Frequency of Cold Waves.—The cold waves that sweep over the country in the rear of cyclonic disturbances are an important feature in the climate of the United States, and the number of rapid changes in temperature depends on their frequency. The amount of these sudden falls of temperature is quite a different matter from the average variability of the temperature just mentioned, and bears somewhat the same relation to it as the absolute range of temperature bears to the average range of temperature. It is not the coldest region, however, that has the greatest number of rapid changes in the temperature. It is rather where the cold of the center of the continent struggles to render abnormally cold the air of the region in the neighborhood of the ocean or other large bodies of water, and where the cyclonic areas occur most frequently. Thus there arises in that region the alternating control of the continental cold and the oceanic warmth. The accompanying chart (Fig. 102) shows roughly the rela-

tive frequency, in the central and eastern United States, of falls of temperature of at least 20° F. in 24 hours. Forty such temperature changes occur in the region of northern Michigan for every five in the region of the southeastern United States.

Dates of Earliest and Latest Frosts.—The times of earliest frost in the fall and latest frost in the spring are of great practical importance.

Average Date of Earliest Killing Frost.—This phenomenon is dependent both on the average temperature,

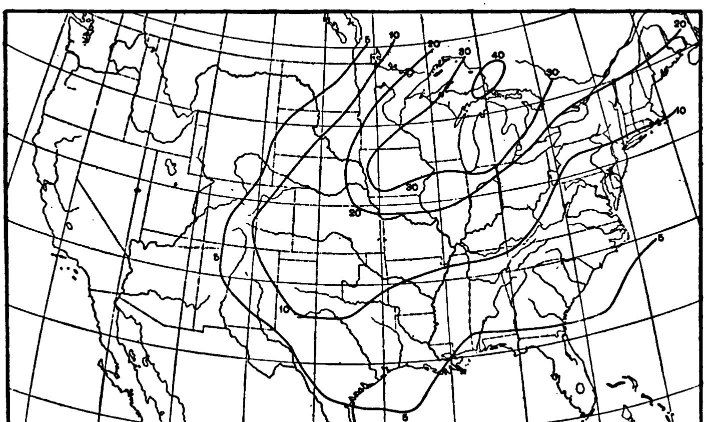


FIG. 102.—RELATIVE FREQUENCY OF FALLS OF TEMPERATURE OF OVER 20° IN 24 HOURS (AFTER RUSSELL).

consequently on the varying altitude of the sun, and on the diurnal range of temperature, and consequently on the cloudiness and distance from oceans. Whenever the average temperature in its autumnal decrease reaches the place where the diurnal range will take the night minimum temperature below freezing, then the first frost must occur; but this regular occurrence is usually anticipated by the ap-

pearance of one of the cold waves just mentioned, in which the temperature fall is far greater than would occur during the regular diurnal change. Frost occurs earliest (Fig. 103) at the Canadian boundary; but the time is retarded, at first gradually, and then more rapidly, with southern progress to the Gulf coast. There is also a retardation from Montana towards the northwest coast of the United States. Frost occurs in the northern part of the United States about Sept. 1; but the time is gradually retarded to Oct. 1 at

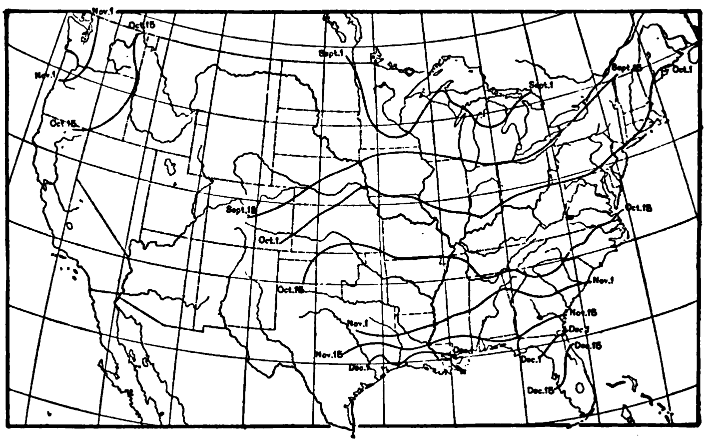


FIG. 103.—AVERAGE DATE OF EARLIEST HARD FROST IN THE UNITED STATES (AFTER GREELY).

about the latitude of the Ohio River valley, and to Nov. 1 in central Mississippi; to Dec. 1 in southern Louisiana, and nearly to Jan. 1 in central Florida (when it occurs at all in the last-named region).

Average Date of Latest Killing Frost.—When with the increase of the average temperatures in the spring (with the altitude of the sun) these become so high that the diurnal change is such as to keep the night minimum

temperature above the freezing point, then ordinarily the frosts will cease; but some temperature falls greater than that due to the diurnal change will very likely occur at a later date, and thus cause a considerable retardation in the time of latest frost. The latest frost occurs (Fig. 104) earliest along the coast of the Gulf of Mexico; but the time is retarded, at first rapidly, but afterwards more slowly, to the Canadian boundary. It occurs about Feb. 1 on the Gulf coast, March 1 in southern Mississippi,

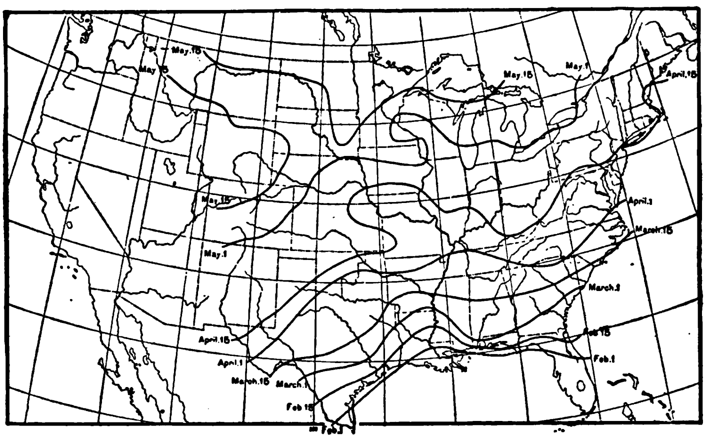


Fig. 104. — Average Date of Latest Hard Frost in the United States (after Greely).

April 1 in northern Tennessee, May 1 in southern Wisconsin, and the latter part of May in North Dakota.

Rainfall.

The rainfall conditions of the United States as a whole are very variable, the total annual amount of precipitation varying from 3 or 4 inches, or less, to over 100 inches.

The accompanying chart (Fig. 105) shows the average

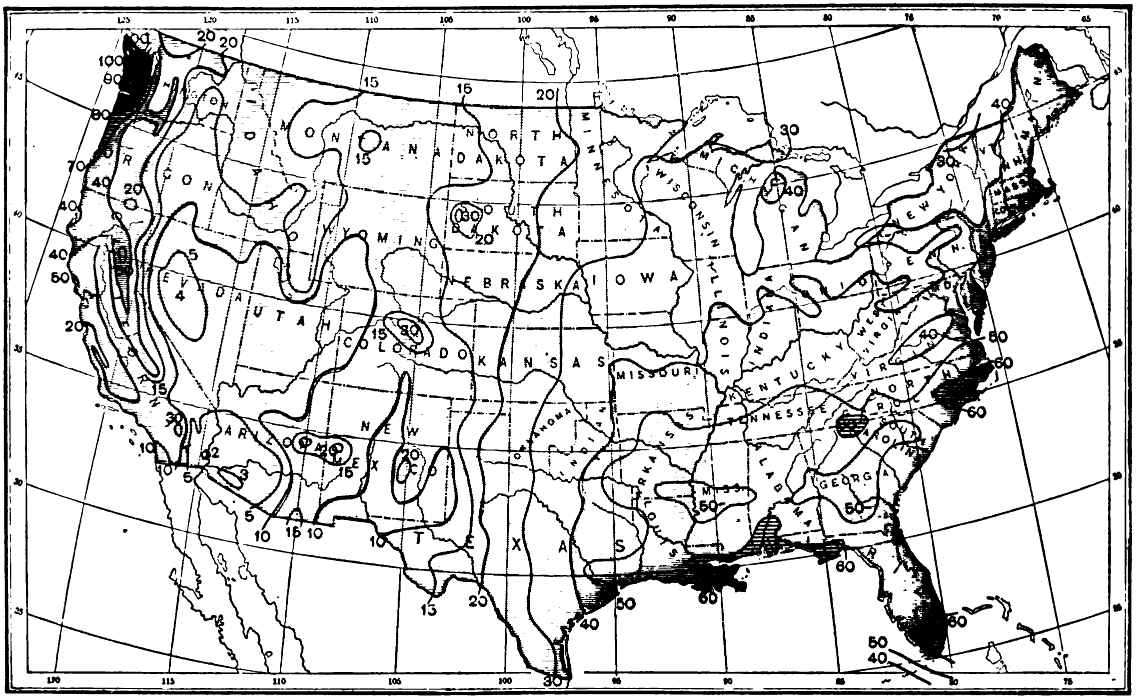


FIG. 105.—AVERAGE ANNUAL RAINFALL IN THE UNITED STATES, IN INCHES (U.S. WEATHER BUREAU).

annual precipitation for the different regions of the whole country.

The Regions of Most Rainfall are: the northwest coast, where in the extreme corner over 100 inches fall, owing to the warm moist air from the Pacific Ocean blowing on to the cooler continent; the northeastern shore of the Gulf of Mexico, where over 60 inches fall, and where a plentiful supply of moisture is found in the southerly winds from the Gulf; the southeast coast of Florida and the extreme eastern portion of the North Carolina coast, where over 70 inches fall, and where a plentiful amount of moisture is furnished by the easterly winds blowing from the Gulf Stream in the Atlantic Ocean.

The Region of Least Rainfall is in the southwestern part of Arizona, where the winter winds blow from the dry interior of the continent, and the summer winds blow from a cooler to a hot region, and thus little rain falls.

It is seen from the map that there is an increase of rainfall with northward progress on the Pacific coast, and a decrease from south to north on the Atlantic coast. In the eastern inland part of the United States there is a decrease from south towards the north; in the central part, in the neighborhood of the 100th meridian, there is no change in a north-and-south direction, but rapid changes in an east-and-west direction; while in the extreme western inland region there is an increase with northward progress. The region separating the western from the eastern type of variation with latitude is almost at the center of the continent.

The wind from the west and southwest blows the moist Pacific Ocean air over the land; and where the land is cooler, as on the northern Pacific coast, an excessive rainfall occurs, but where the land is warmer, as on the south-

ern coast, but slight precipitation occurs. Moreover, the sea wind at the south is frequently but a day wind, and is not as permanent as in the north.

The air flowing into the northwestern United States is deprived of most of its moisture before it has gone far inland, so that the regions to the east have dry air and little precipitation until moist air is supplied by winds from the Hudson Bay, the Great Lakes, the Atlantic Ocean, or the Gulf of Mexico. And the northwest and northern winds carry this dry air far to the south, almost to the Gulf of Mexico, making the whole region traversed one of deficient rainfall. But from the region where the southwest winds begin to carry the moist air from the Gulf of Mexico up into the Mississippi Valley, the rainfall increases again rapidly over the whole eastern United States. The excessive precipitation of the southeastern United States is due to the prevalence of warm oceanic winds.

The whole United States is divided into three great distinct regions as regards rainfall.

The first region, to the east of the 97th meridian (which passes through central Kansas), has an average rainfall of over 35 inches; and while there is a decrease from nearly 60 to 35 inches with northward progress, yet, on the whole, the distribution of the amount of rainfall is remarkably equable for such an extended region.

The second region, that to the west of the 97th meridian (central Kansas), has at first, for the next 5° of longitude, a rapid decrease from 35 to 15 inches of rainfall at the western border of Kansas; but from this (102d meridian) westward there is little change in the rainfall north of latitude 42° (the southern boundary of Wyoming) to the region of the northwest coast, where the increase becomes very rapid towards the coast. From about the 100th meridian westward, and south of latitude 40°, there is a slow, somewhat variable decrease from 20 inches to a minimum rainfall of from 2 to 5 inches in the desert region to the east of the Sierra Nevada Mountains, from which there is a slight increase to the Pacific coast. The rapid change in rainfall in

the region just east of the Rocky Mountains is due to the fact that just there the moisture begins to be received from the Gulf of Mexico on the south, and Hudson Bay and the Great Lakes on the north.

The third region is on the northern half of the Pacific slope, to the west of the Cascade Mountains. Here the rainfall increases rapidly (from 25 inches) from northern California to the northern boundary of Washington, and from the Cascade Mountains on the east to the ocean on the west.

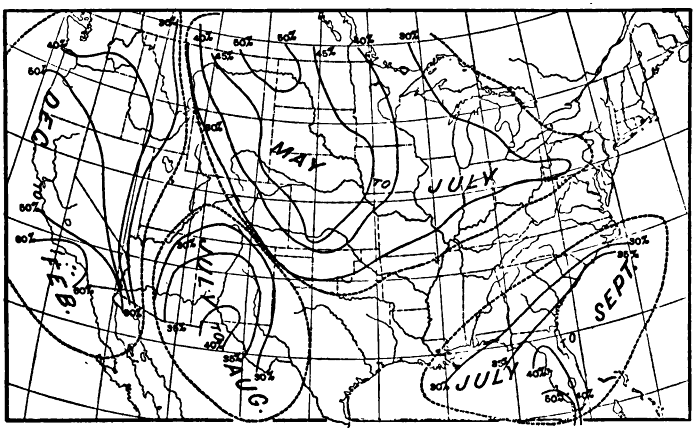


FIG. 106. — PERCENTAGE OF TOTAL RAINFALL IN THE RAINY SEASON. MONTHS OF RAINY SEASON. (U.S. WEATHER BUREAU.)

Periods of Rainfall.—The accompanying chart (Fig. 106) shows the months of greatest rainfall, and the percentage of the total annual rainfall which falls during those months.

The dotted lines are drawn merely to roughly separate the various regions, and to show that the months and lines of percentages inclosed by each belong together. The regions between these must form transition zones from one system to another.

In most of the United States the period of greatest rain is that usual for continental localities, except on the Pacific coast, where there is a winter rainy season.

In the winter time, the warm, moisture-laden winds of the Pacific are blown on to the colder west coast of the United States, and precipitation takes place. In the summer time, however, this moisture is rendered still warmer, and thus farther removed from the temperature of condensation.

Much of the summer rainfall of the eastern United States is of local character, and accompanies thunderstorms.

Rainfall Types. — The classification of the rainfall for the United States by regions is about as follows, when the two types of a single and a double period for the annual amount of rainfall are made the basis for classification: —

1. The *Pacific type of rainfall* is characterized by heavy precipitation during midwinter, when the west and southwest winds blow from the ocean; and an almost total absence of rain in the late summer, when the winds blow from the ocean to the warmer land, or when the north winds blow from the dry continent. This type is found in British Columbia, Washington, Idaho, Oregon, California, Nevada, and western Utah.

2. The *Mexican type of rainfall* is characterized by heavy precipitation after the summer solstice, when the southeast winds blow in the moist Gulf air; and a very dry period after the vernal equinox. This type is found in Texas, New Mexico, and most of Mexico. In Arizona the type is complex, resulting from the meeting of the Pacific and Mexican types.

3. The *Missouri type of rainfall* is characterized by light winter rains, because the prevailing north and northwest winds bring little moisture; and heavy rains due to the south and southeast winds blowing the moisture from the Gulf in the late spring and early summer, when most of the annual rainfall occurs. This type occurs over the slopes of the Arkansas, Missouri, and Upper Mississippi rivers, and Lakes Superior and Michigan, and includes the States of Montana, the Dakotas, Minnesota, Nebraska, Kansas, Iowa, Missouri, Wisconsin, Illinois, the Territory of Oklahoma, with parts of Arkansas, Texas, Michigan, Indiana, and Indian Territory. This covers the greatest area of any; and the time of occurrence of the rainfall is most favorable to agriculture, considering the total amount of annual rainfall.

The eastern part of Texas has a complex rainfall, resulting from the meeting of the Mexican and Missouri types.

4. The *Tennessee type of rainfall* has the heaviest rains in the latter part of the winter or the early spring, when moist south winds from the Gulf meet the cold north winds from the interior; and the least rain in mid-autumn, when the cool, dry wind blows from the north and northeast. This occurs in Tennessee, Arkansas, Mississippi, eastern Kentucky, western Georgia, Alabama, and Louisiana.

5. The *Atlantic type of rainfall* has a somewhat uniform distribution of rain throughout the year, and it covers the Atlantic coast States except New England.

6. The *St. Lawrence type of rainfall* has heavy rains in the late summer and autumn months, due to the meeting of warm, moist south, and cool north winds; and a scarcity of rain in the spring, due to the dry winds from the west and north. It covers the St. Lawrence valley.

In New England the Atlantic and St. Lawrence types meet, and cause late summer and late fall maxima, and irregular minima in early summer and spring or early fall.

Agriculture Rains. — The rainfall during the late spring and early summer months in the United States is very favorable for agriculture east of the Rocky Mountains, and for very limited regions to the west. A well-distributed rainfall of over two inches per month is found east of about the 100th meridian, in April; east of the 115th meridian (central Idaho) in the northern United States, and east of the 102d meridian (western Texas) in the southern United States, in May; and east of the 115th meridian at the north (central Montana), the 102d meridian at the center (eastern Colorado), and the 100th meridian at the south (central Texas), in June. It is to these widely distributed sufficient rainfalls that the agricultural prosperity of the United States is largely due. In most of the central and eastern United States the rainfall is between two and four inches per month. Where

it is over six inches per month in limited regions, mostly near the Gulf coast, the excess is due to a few heavy rains.

Excessive Rains are those which are injurious to plant growth, and which cause damaging floods. When the rainfall exceeds two inches in a day, or ten inches in a month, it becomes excessive. Such rains are most frequent on the North Pacific coast from January to March, on the South Atlantic coast in summer, on the Gulf of Mexico in the spring, and in the Tennessee and Kentucky region from the late winter to early summer. Occasionally these excessive rainfalls occur north of the Potomac River, chiefly in the late summer or early fall.

An *excessive monthly rainfall* of nearly 42 inches has occurred in northern California, one of 37 inches in Louisiana, one of over 28 inches in North Carolina, and one of over 22 inches in New Jersey.

The *greatest daily rainfalls* (in 24 hours) do not usually exceed 10 inches, with occasionally double that amount during two successive days. The chart, Fig. 107, shows the greatest daily rainfall in various regions of the United States.

On June 15 and 16, 1886, and within a period of 24 hours, nearly 21.5 inches of rain fell at Alexandria, La. On Feb. 11–13, 1886, there occurred a remarkable storm in New England, in which precipitation was excessive over a large extent of territory. The amount was 12.4 inches at Canton, Conn.; and it averaged more than 7 inches over 1,500 square miles, and more than 5 inches over 5,000 square miles.

Rain frequently falls at the much more rapid rate of from 5 to 18 inches per hour during short intervals of time. Such rainfalls usually last from 2 or 3 to 15 minutes, and rarely for so long as half an hour. In Galveston, Tex., in June, 1871, there was a rainfall of nearly 4 inches in 14 minutes; but usually the amount during these intense

rainfalls is about 2 inches. We do not know much about the amount of rainfall in the cloud-bursts, except from the overflowing of creeks and the damage which is done.

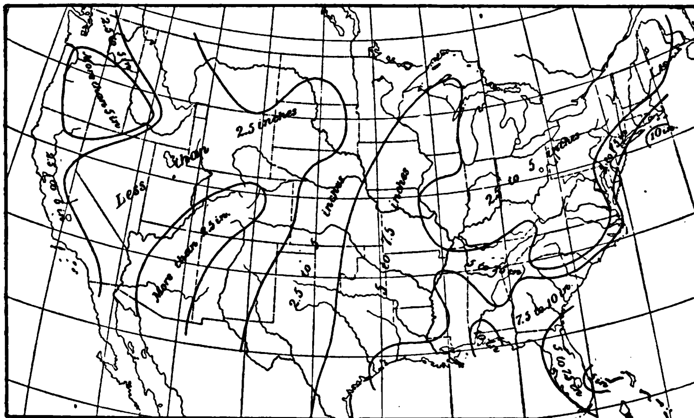


FIG. 107.—HEAVIEST DAILY RAINFALLS (U.S. WEATHER BUREAU).

Snowfall in the United States.—Snow falls with greater or less frequency in almost all parts of the United States. Since snowfall depends on the temperature, snow occurs in very low latitudes if the altitude is sufficiently great; but it is almost unknown on the Florida peninsula and on the California coast below the 35th parallel. Taking the country as a whole, snow seldom remains unmelted on the ground south of about latitude 33° at low altitudes; and on the eastern coast this limiting latitude is about 31°, while on the western coast it is about 37°.

The region of greatest snowfall would naturally be that in which the greatest amount of precipitation occurs in winter, but it is also necessary that the temperature be low enough to convert the moisture into snow. The total average annual amount of snowfall in inches, for the

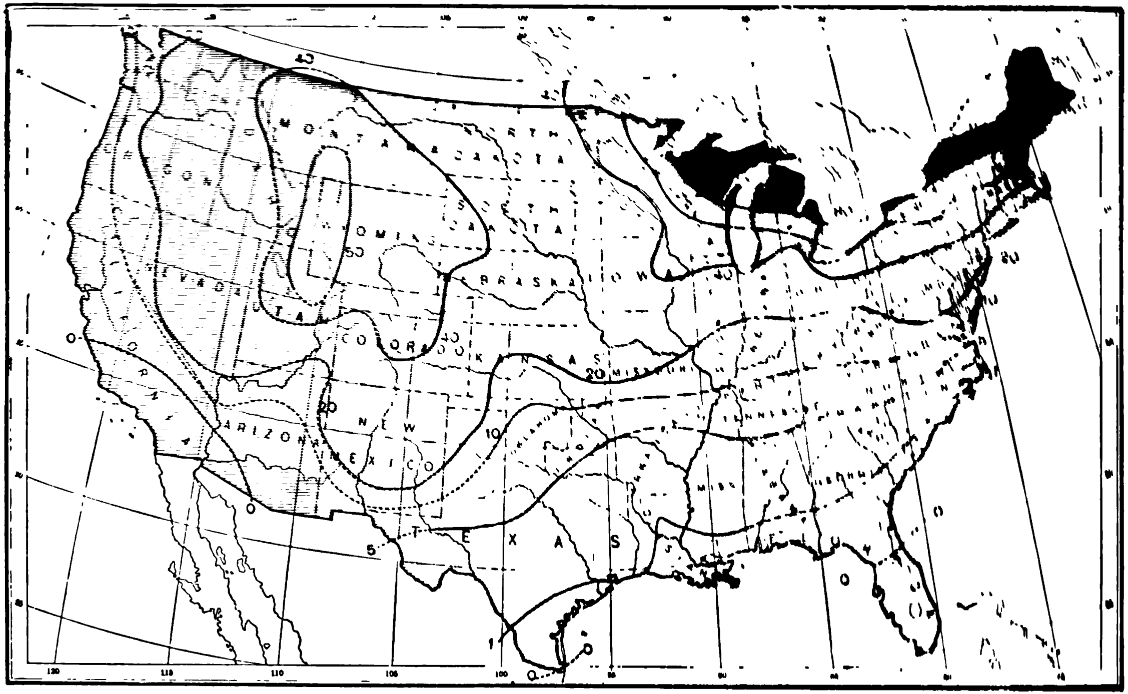


FIG. 108. — AVERAGE ANNUAL SNOWFALL IN THE UNITED STATES (DEPTH IN INCHES).

7-year period 1884–91, is shown on the accompanying chart (Fig. 108).

The lines connect the regions of equal snowfall, and the attached figures indicate the amounts of snowfall in inches. The snowfalls represented on this chart refer to the low lands and plains in which the larger towns are situated. Where there are mountains, the present figures might have to be increased even several fold in order to represent the snowfall at higher altitudes than their bases.

In general, for the low lands, the region of greatest snowfall is from northern Michigan (where it is 130 inches) eastward along the Canadian frontier. To the southward of this region there is a rapid decrease to 20 inches at about the latitude of St. Louis, and then the decrease is very slow until the practical disappearance of snow near the Gulf coast. On the Atlantic coast the snowfall is 60 inches at the north, and diminishes, at first rapidly and then more gradually, to the limit of practical disappearance on the Carolina coast. To the west of the region of maximum snowfall in northern Michigan, there is a diminution to nearly 30 inches in the Missouri valley, and then an increase to 50 inches on the high plateaus from Utah to Montana, with a decrease farther west towards the Pacific Ocean. In the whole southwest United States the snowfall is slight, except on the mountains. On the Pacific coast the snowfall is less than 10 inches at the north, and decreases to less than 1 inch at San Francisco; while south of San Francisco snow seldom falls. Along the Mississippi River there is a decrease from 40 inches at the extreme north, to 20 inches at St. Louis, and 2 inches at Vicksburg. The greatest recorded annual amounts of snowfall are in eastern California and western Nevada, where they reach several hundred inches. At

Summit, Cal., on the Central Pacific Railroad, the snowfall averaged 378 inches per year during a period of ten years.

For the greater portion of the United States it may be said that in the northern half the snowfall exceeds 20 inches, and in the southern half is less than 20 inches.

Frequency of Precipitation.—The distribution of the number of days with precipitation (rainy and snowy days) follows in part the amounts of precipitation, and in part

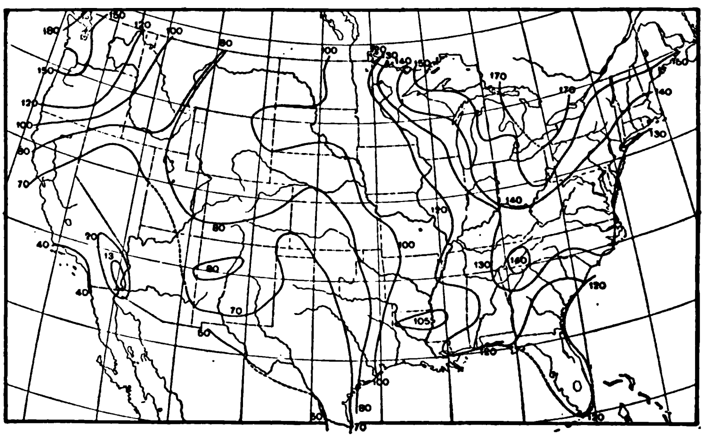


FIG. 109.—AVERAGE NUMBER OF DAYS DURING THE YEAR WITH PRECIPITATION (U.S. WEATHER BUREAU).

the distribution through the year. For the whole year, along about the central meridian of the United States, there is precipitation on about 100 days (Fig. 109). To the eastward there is an increase towards the Atlantic Ocean, with a local maximum over the lower Great Lakes. To the westward there is an increase towards the Pacific Ocean in the northern part, and a decrease in the southern part.

In the western United States, where the seasons of rainfall are strongly marked, the number of rainy days is

greatest in the region of greatest rainfall, and least in the region of least rainfall. In the eastern United States, however, the region of greatest number of days is at the north, where the rainfall is more equably distributed over the whole year, and where the passage of barometric minima is most frequent.

The Greatest Number of Consecutive Days with Precipitation is shown on the chart, Fig. 110. In general,

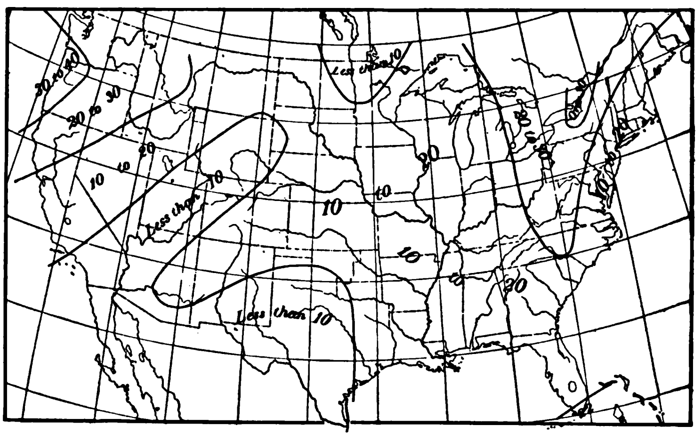


FIG. 110. — GREATEST NUMBER OF CONSECUTIVE DAYS WITH PRECIPITATION (U.S. WEATHER BUREAU).

the regions of greatest rainfall have the greatest number of consecutive days with precipitation. For the main part of the United States, the number of days is from 10 to 20; but on the northwestern coast, where there is such a pronounced rainy season, with steady winds from the ocean, there is an increase to nearly 40 days; while in the northeastern United States in the St. Lawrence valley, where the cyclonic areas are most frequent, the number of days increases to over 30.

The Greatest Number of Consecutive Days without Precipitation is of importance as signifying the duration of *droughts*, and is shown on the chart, Fig. 111. The more equable the distribution of precipitation through the year, the less the number of consecutive days without it. In nearly the whole of the eastern half of the United States, the greatest period of uninterrupted drought is from 15 to 30 days, except in the extreme southeastern part, where it varies from 30 to about 50 days. There is an increase in

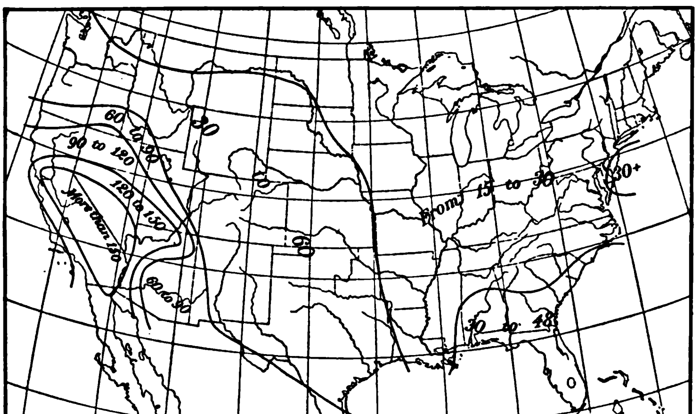


FIG. 111. — GREATEST NUMBER OF DAYS WITHOUT PRECIPITATION (DROUGHT) (U.S. WEATHER BUREAU).

the number of days from the center of the United States towards the northwest, to from 30 to 60 days, and towards the west and southwest to over 150 days.

Absolute Humidity. — The absolute amount of water contained in the air is shown on the charts, Figs. 112, 113, which give the number of grains contained in each cubic foot of surface air for the various regions of the United States. Since the possible amount of water in-

creases with the increase of the temperature, the distribution follows somewhat after that of the temperature, and is least in winter and greatest in summer; but the distance and accessibility of the supply of moisture, and the direction of the wind, also enter into this distribution.

In the Winter (Fig. 112) there is a region of least moisture in the region of lowest temperature in the north central United States; and there is an increase towards the east, south, and west, in the direction of increase of

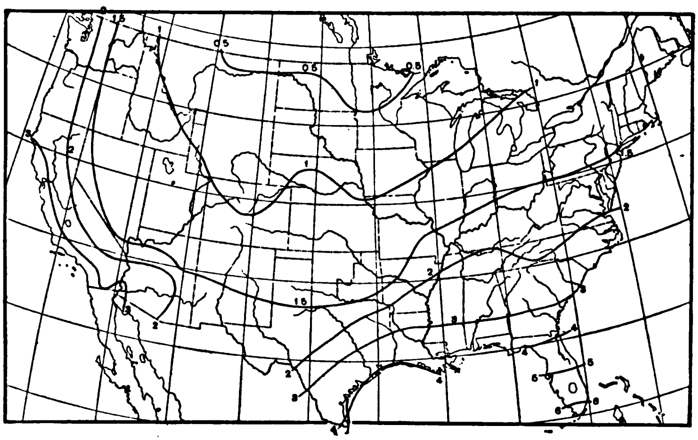


FIG. 112. — AVERAGE ABSOLUTE HUMIDITY IN MIDWINTER, JANUARY (AFTER GREELY).

temperature, and roughly in proportion to this increase of temperature.

In the Summer (Fig. 113) the humidities are much increased over those of winter, and the increase is proportionally the greater where there is the greater annual range of temperature, although the increase in grains is nearly the same in all parts. The region of least moisture is now shifted to the high plateau west of the Rocky Mountains, where, with the region to the west of it, the

average temperatures are lowest. From thence there is an increase towards the higher temperatures to the east, southeast, and south. In the southwestern part of the United States the moisture would undoubtedly become

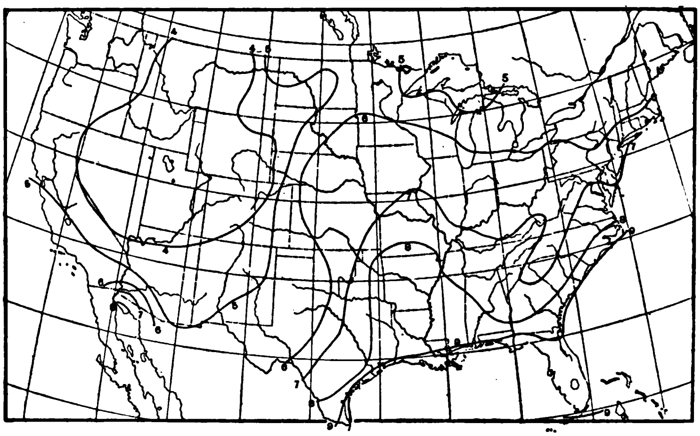


FIG. 113. — AVERAGE ABSOLUTE HUMIDITY IN MIDSUMMER, JULY (AFTER GREELY).

greater if it were not for the mountain ranges cutting off the winds, and the unfavorable wind direction with regard to the supply of moisture.

The Relative Humidity (Fig. 114), depending as it does on the absolute moisture and the height of the temperature above that necessary for condensation, is least in the southwestern United States, where the amount of moisture is low and the temperature is high. There is an increase in all directions from this region to the limits of the United States. This increase is fairly symmetrical, although not with the same rapidity on all sides. The relative humidity, while lowest in summer and highest in winter, is so irregularly distributed throughout the year, that midwinter and midsummer charts are not given.

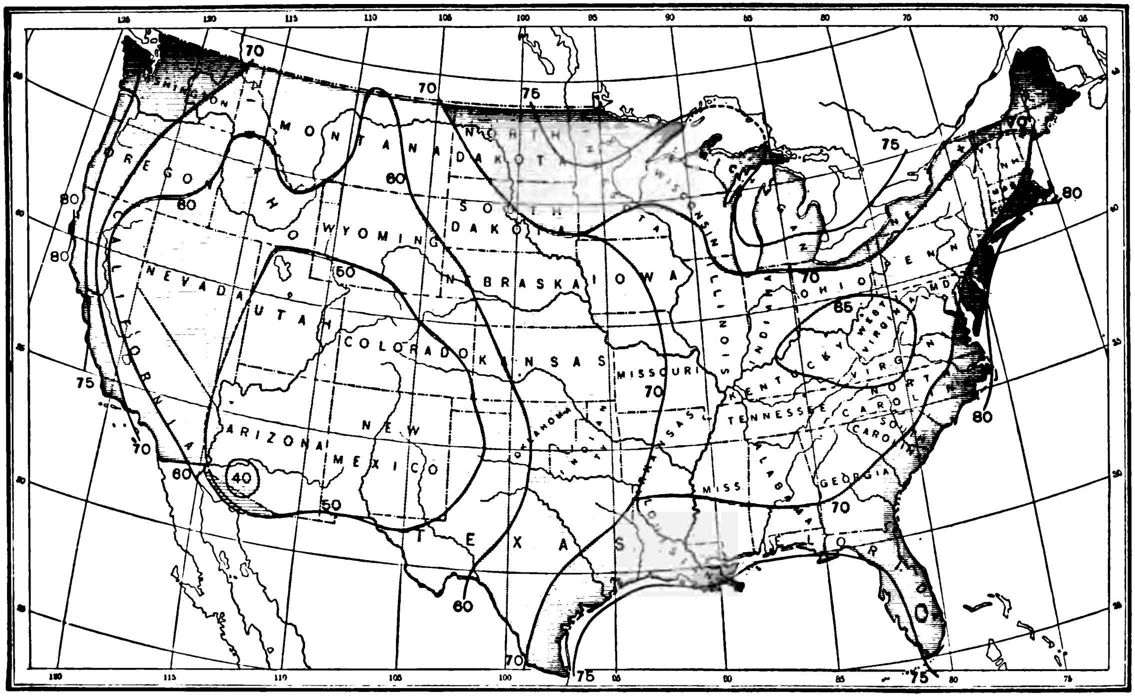


Fig. 114. -- Average Annual Relative Humidity in the United States.

The average relative humidity for the year (Fig. 114) is distributed over the United States about as follows: On the Atlantic and Pacific coasts there is a relative humidity of about 80% on the northern coast, which decreases to a little below 75% on the southern coast; on the coasts of the Great Lakes and the Gulf of Mexico it is about 75%; in the low eastern half of the inland United States it is about 70%, except in the upper Ohio River valley and western Virginia, where it decreases to 65%; in the elevated western half of the inland United States it is below 60%, and decreases gradually to 40% in southwest Arizona.

Degree of Cloudiness. — The observations of the degree of cloudiness are made on a scale of from 0 for clear weather to 10 for an entirely overcast sky, but the average results are as usual expressed in percentage of complete cloudiness. Thus 50% of cloud signifies that the sky is half covered with cloud.

The degree of cloudiness, which depends on the relative humidity, varies with the season of the year somewhat irregularly, but is in general greatest in winter and least in summer; but the maximum occurs in the spring months in the Rocky Mountain region, and over the Great Plains at the center of the continent.

Cloudiness in January (Fig. 115). — In the extreme northwest of the United States there is a cloudiness of over 70%, which decreases towards the south to 40% on the south Pacific coast, towards the southeast to between 30% and 40% for nearly the whole of the southwestern part of the United States, and towards the east nearly to 40% on the northern Great Plains. At nearly the longitudinal center of the continent there is a cloudiness of about 50%, and to the eastward of this there is a quite rapid increase in the

northern part to over 70% on the Lower Lake region, with a decrease to less than 60% on the North Atlantic coast; while in the southeastern part of the United States there is at first an increase to about 60%, and then a slight decrease again as the coast is approached. In Florida the percentage falls below 50%, except on the eastern coast.

The central line of the region of least cloudiness passes from southern Arizona in a northeasterly direction near the northwest corner of New Mexico, the western part of Colorado, the southwest corner of

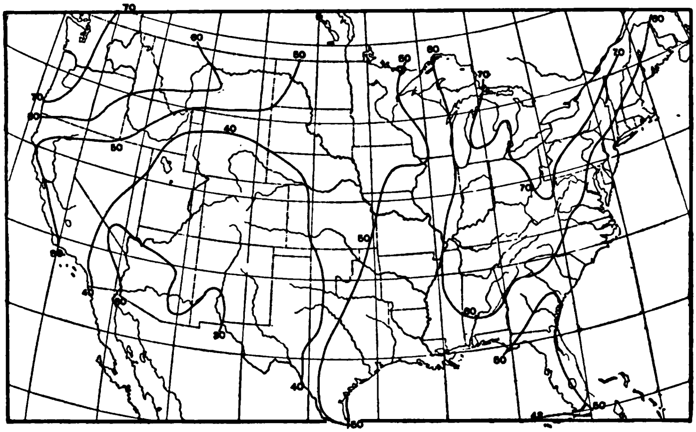


FIG. 115.—AVERAGE CLOUDINESS IN MIDWINTER, JANUARY (AFTER GREELY). (*Scale: 0 = clear; 100 = completely cloudy.*)

South Dakota, and the northeast corner of North Dakota. To the east and west of this line there is a very rapid increase in the cloudiness in the northern part, and very much smaller increase in the southern part, of the country.

Cloudiness in August (Fig. 116).—The cloudiness in August is least in the Sierra Nevada region, where it is less than 10%, and is greatest on the southern Pacific coast, the Atlantic coast, and in a small region in north-

western New Mexico, where it varies from 45 % to 50 % of total cloudiness.

In the northern half of the United States there is an uninterrupted increase in the cloudiness from 20 % in the western part (Idaho) to about 45 % in the region of the Great Lakes, whence it varies little to the Atlantic coast. The increase is slow west of the Rocky Mountains, but east of them is much more rapid. In the southern half of the United States there is a cloudiness of 45 % on the

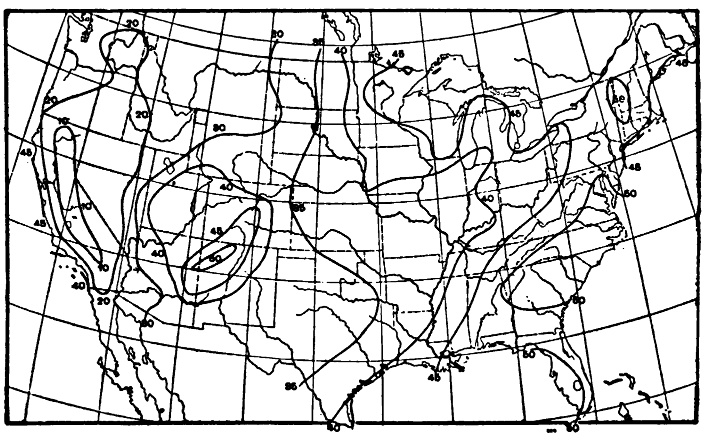


FIG. 116. — AVERAGE CLOUDINESS IN MIDSUMMER, AUGUST (AFTER GREELY).
(*Scale : 0 = clear ; 100 = completely cloudy.*)

Pacific coast, and a rapid decrease to 10 % in the near inland (central California), a local increase to 45 % or 50 % in New Mexico and Colorado, then a decrease to 35 % at the 100th meridian, and thence eastward a steady increase to 50 % on the eastern coast.

At the meridional center of the continent, the cloudiness is from 35 % to 40 %, and in general decreases towards the west, and increases towards the east, by about 15 %.

Winds.

Average Wind Velocity for the Year, in Miles per Hour (Fig. 117). — If the average amount of wind which blows during the entire year were evenly distributed through all the hours of the year, it would make a uniform velocity of about 13 miles per hour on the North Atlantic coast, 14 on the central coast, and probably decrease to 11 or 12 on the southern coast. On the shores of the Great Lakes, it

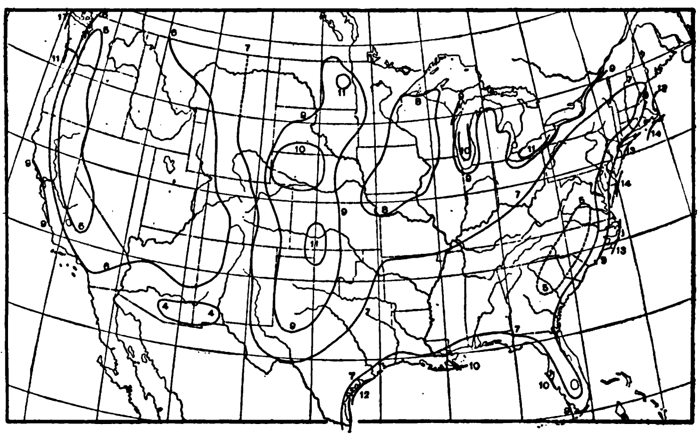


FIG. 117. — AVERAGE WIND VELOCITIES FOR THE YEAR (IN MILES PER HOUR).

would be about 11 miles per hour; and on the Gulf coast, in the western part about 12, but in the eastern part about 10, miles per hour. On the Pacific coast, a velocity of 11 miles per hour at the north decreases to probably 8 miles or less per hour at the south. On the Great Plains, extending from northern Dakota to Texas, the velocities would be about 10 miles per hour, with a maximum of 11 in central Kansas. To the west of about the 100th

meridian there is a decrease to 6 miles per hour on the Pacific slope along the region of central California, Oregon, and Washington. Towards the east from the 100th meridian there is a decrease to 8 miles per hour in the northeastern part of the United States (except directly along the Lakes), and to less than 7 miles per hour for most of the region south of the Ohio and east of the Mississippi River. The velocities are thus the greatest on the coasts and treeless prairies, and least inland and in the mountainous and wooded regions.

Month of Maximum and of Minimum Wind.—The month of maximum wind varies somewhat in the different regions of the United States, but usually occurs in March in the eastern half, and in April or a little later in the western half, of the country. The time of minimum is not quite so irregular, and occurs usually in August.

At low altitudes in the eastern United States, away from the coast, and as far west as eastern Kansas, the maximum wind occurs in March; but where the stations have a greater altitude, or are directly on the coast, it occurs in an earlier month, perhaps as early as December, or even November. In the region to the west of the 100th meridian (central Kansas), and extending to the Pacific slope, the month of maximum is irregular, and at moderate altitudes occurs in April or sometimes in May or June; but at very high altitudes there is a return to the beginning of the year, which was noticed in the east. On the Pacific slope the months are not the same in different localities. On the southern coast the maximum occurs in April, and the minimum in November; on the central coast the maximum is in July, and the minimum in November; on the northern coast the maximum is in January.

Amount of Maximum and Minimum Monthly Winds. — The average amount of wind for the months of greatest and least wind is about 17 miles per hour for the greatest, and 9 for the least, on the North Atlantic coast; probably 14 or 15 for the greatest, and 9 to 12 for the least, on the South Atlantic coast; about 11 or 12 for the greatest, and 6 for the least, on the South Pacific coast; and 16 for the greatest, and 7 for the least, on the North Pacific coast. On the Great Lakes the average velocity is about 14 miles per hour for the greatest, and 9 for the month of least wind; while on the Gulf coast it is 15 for the greatest and 10 for the least at the western part (Texas), and only 11 for the greatest and 8 for the least at the eastern part (Florida). These numbers apply to the immediate shore; and even a single mile inland the velocities are much reduced, probably as much as 33 % near the ground even if the intervening land is level.

Away from the immediate coast the average monthly wind velocities reach about 9 miles per hour for the greatest and 6 for the least in the southeastern part of the United States; 11 for the greatest and 6 for the least in the northeastern part; but increase with westward progress to 13 miles per hour for the greatest and 9 for the least on the Great Plains from North Dakota to southern Texas, and decrease still farther to the west to 8 for the greatest and 6 for the least in the southwest United States, and 11 miles per hour for the greatest and 6 for the least in the northwest United States.

The difference between the average amount of wind in the windiest and calmest months of the year, expressed in percentage of the average wind, is about as follows: On the North Atlantic coast region, from 50 % to 60 %; on the central coast, 40 %; and on the southern coast, 30 %; on the Upper Lakes, 30 %, and on the Lower Lakes, 40 %; on

the Gulf of Mexico coast, 40 %; on the Pacific coast, 70 % at the north, 60 % at the center, and probably not over 40 % at the south. Inland, east of the Rocky Mountains, there is an increase from 30 % to 50 % from the northern boundary of the United States towards the south, and towards the southeast as far as the western boundary of North Carolina, but still farther to the southeast there is a decrease to 30 %. To the west of the Great Plains there is an increase to about 60 % along the Rocky Mountain region, but it decreases again to only 20 % in central Nevada, whence it gradually increases again towards the coast.

A marked peculiarity of this feature is that the same percentages found on the coast sometimes extend far inland; and consequently these ratios expressed in percentages hold good for the region in which they occur, irrespective of a land or water surface.

Maximum Wind Velocities. — The maximum velocities attained in individual wind storms are greatest in regions of tornadoes and severe thunder squalls, but these winds are of short duration. Strong winds which blow for a considerable length of time, say for over five minutes, have velocities of from 50 to 70 miles per hour inland, and from 70 to 90 miles per hour on the coasts of the United States. Inland the velocities are less at the south than farther north; but on the southern Atlantic and Gulf coasts the hurricane winds are as strong as those experienced on the northern coast. On the Pacific coast the velocities increase from less than 40 miles per hour at the south to over 90 at the north. The southwest United States is the region of least maximum winds.

Wind Direction. — The direction of the wind is in general towards the east with the primary air currents, but it suffers interruption from monsoon effects and the passage across the country of areas of high and low air pressure. The average wind directions for any length of time should conform to the distribution of the air pressure during that time; and this distribution, in extreme seasons,

over the United States is probably about as given in Figs. 118, 119; but there is considerable uncertainty about the pressures in the elevated western region.

There is a strong tendency for winds of a monsoon character to arise, especially in the great central region extending from the Gulf of Mexico to Canada; but the frequency and magnitude of the cyclonic areas prevent these from becoming so well marked as they would be in lower lati-

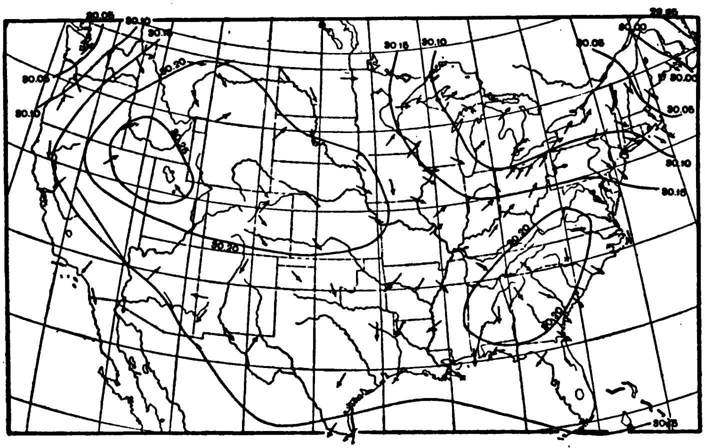


FIG. 118.—NORMAL AIR PRESSURE AT SEA LEVEL, AND NORMAL WIND DIRECTION, FOR JANUARY (U.S. WEATHER BUREAU AND HAZEN).

tudes. We can consider in detail only the midwinter month of January and the midsummer month of July.

The Average Wind Direction for January is shown on Fig. 118. The surface wind directions, in the greater part of the United States, are mainly the result of the anticyclonal motion around the region of high air pressure over the interior of the continent, acting on the general current towards the east. The low pressure at the northeast strongly affects the New England winds. Along the 40th

parallel the wind direction is towards the east all the way across the continent, except on the high plateau of Utah, where it is northerly. To the north of this parallel we find in the extreme northwest in Washington a wind towards the west; in Idaho it swerves around towards the north, in Montana it has turned towards the east, and in the Dakotas towards the southeast; and more to the eastward, in the Great Lake Region, there is a tendency for the winds to turn in a direction to the north of east, but in New England there is a bend again towards the southeast, owing to the area of low pressure at the northeast. Between the 40th and 30th parallels there is a wind towards the south in the Western States, towards the southwest in the Central States, and towards the southeast in the Eastern States. South of the 30th parallel the wind is towards the west in the eastern part, but swerves around towards the south in the Western States.

The Average Wind Direction for July is shown in Fig. 119. The area of low pressure probably centering over Arizona and Utah controls the direction of the surface wind for the region west of the 100th meridian, except at the extreme north, where the wind direction is towards the east; so that in the western United States there is a wind towards the south at the north-central part, towards the west at the eastern part, towards the east at the western part, while at the southern part the wind is towards north and east. The area of low pressure in the northeastern United States (and Lower Canada) gives the easterly direction to the wind in that region. The area of moderately high pressure over the southeastern United States furnishes winds blowing outwards towards the areas of low pressures to the west and to the north. In the southeastern part of the United States the winds are towards

the north; in the eastern section of this region they incline towards the east of north (towards the area of low pressure in the northeastern United States), and in the western part towards the west of north (towards the area of low pressure in the western United States).

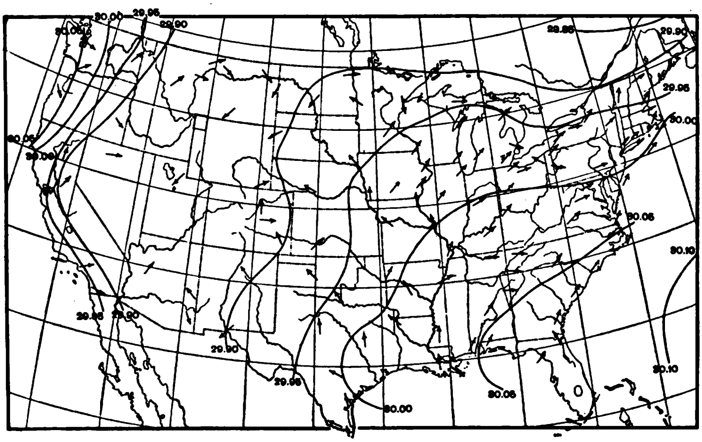


FIG. 119.—NORMAL AIR PRESSURE AT SEA LEVEL, AND NORMAL WIND DIRECTION, FOR JULY (U.S. WEATHER BUREAU AND HAZEN).

Rain-bearing Winds.—The direction of winds most likely to be followed by precipitation, for the whole year, is given on the chart, Fig. 120, the limits being marked by the arrows. These wind directions depend a good deal on the direction of the water supply, and the nature of the intervening country; and they vary somewhat from month to month. Such charts are very useful in making predictions of rain in weather forecasting.

Destructive Storms in the United States.—The greatest number of destructive storms, which are usually of a thunder-squall or tornadic character, occurs in the months of June and July. They occur more frequently in the spring

than in the fall; and there is a retardation with the advance of the season and with increase of latitude. Thus, in the southern part of the United States, tornadoes are frequent in the spring, but in the northern part they are most frequent when summer is well advanced. They seldom occur west of the 100th meridian, since the air farther west is

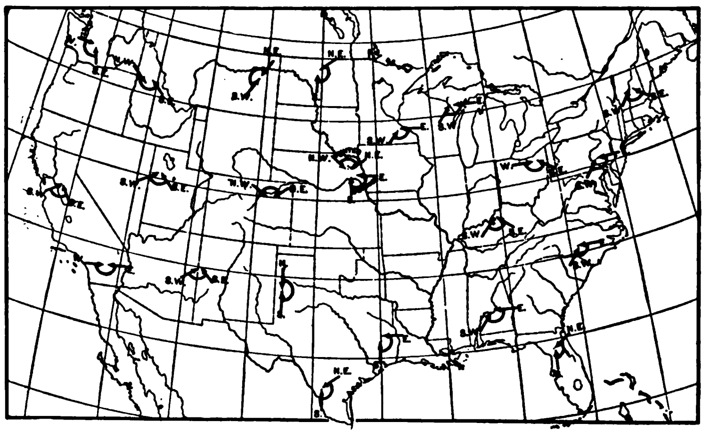


FIG. 120. — WIND DIRECTION MOST LIKELY TO BE FOLLOWED BY RAIN.

too dry for their formation. The greatest number of destructive storms occur in Illinois, Iowa, and Pennsylvania, and the least number in California.

The chart, Fig. 121, shows the average annual number of thunderstorms in the United States, and the months of their greatest frequency. It is seen that for the great central region, including the Mississippi River and Ohio River valleys, the month of greatest frequency is June; but there is a retardation of about two months with approach to the eastern and southern coasts, and also with the progress into the region of deficient moisture in the western part of the United States.

The region of greatest frequency of thunderstorms is in the southeastern part of the United States near the coast, where, on the average, 40 occur during the year. There is a decrease in all directions, and the number is reduced to about 20 at about the 105th meridian on the west, at the latitude of the northern boundary of the United States on

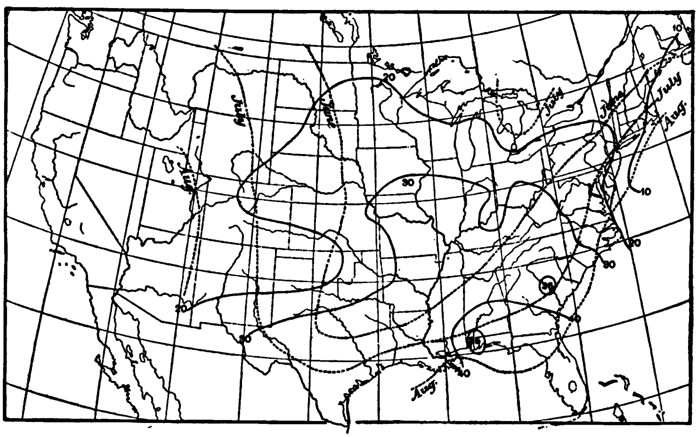


FIG. 121.—MONTHS OF MAXIMUM FREQUENCY, AND AVERAGE ANNUAL NUMBER, OF THUNDERSTORMS (U.S. WEATHER BUREAU).

the north, and at the Virginia coast on the northeast. On the New England coast only 10 occur.

We have now touched on the main features of the climate of the United States. A study of charts will reveal many details which it has not been possible to mention specifically, and the pupil is recommended to make frequent study of these maps, and thus become familiar with the characteristics of the climate in the various sections of our country.

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Waldo, F.
Elementary meteorology.

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